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COMMENTS ON SOME EASTERN ADIRONDACK PROBLEMS

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ABSTRACT

Several controversial points in the geology of the Adirondack anorthosite are briefly discussed, and some new observations are added. Balk's postulated evolutionary series, leading from gabbroic anorthosite to large, spherical gabbros, needs revision in view of Buddington's proof that anorthosite is cut by discordant gabbro dikes. The possible derivation of gabbroic anorthosite from Grenville skarn is discussed. A certain amount of pyroxene, and perhaps even of andesine, has been precipitated by volatiles. Accumulations of mafic minerals, therefore, need careful examination to establish their origin.

A rather prolonged stage during which diminishing flowage and initial fracturing overlapped in place and time is thought to explain, to some extent, the observed occurrence of mineral-veneered short fractures with, or without, marginal flexures in the surrounding anorthosite. The origin of ill-defined siliceous veins and of the rare anorthosite dikes may be related to this condition of the anorthosite.

The evidence for a chilled border of the anorthosite is judged to be inconclusive. Some features bearing on the shape of the anorthosite massif are reviewed, and interpretations of syenite dikes in anorthosite are briefly discussed.

INTRODUCTION

The recent memoir by Dr. A. F. Buddington on the geology of the Adirondack pre-Cambrian rocks[1] has greatly advanced our knowledge of this unusually complicated and interesting area. Probably no two geologists who have worked in the region agree on all problems, and thus this district affords opportunity for exceptionally stimulating field work; an explanation or theory that may seem competent to explain one group of features is likely to be challenged sooner or later by alternative explanations from other quarters, based on other features. Very careful field work and considerable caution in interpretation are therefore necessary.

Some years ago, the writer spent a number of field seasons in the eastern Adirondacks studying some of the structural features of the great massif of anorthosite. The principal results were published in 1931.[2] As was to be expected, discrepancies in interpretations and observations have arisen. In 1942, after the completion of a prolonged project in New England, the writer hoped to reinvestigate in detail some of the critical problems. Five weeks were spent in the eastern Adirondacks in the summer of 1942, but subsequent war work has prevented an immediate continuation of the studies. For this reason a few comments and observations are offered at this time which may help to clarify some questions

[1] "Adirondack Igneous Rocks and Their Metamorphism," *Geol. Soc. Amer., Memoir 7* (1939), pp. 1–354.

[2] Robert Balk, "Structural Geology of the Adirondack Anorthosite," *Min. u. petrog. Mitt.*, Vol. XLI, Heft 3–6 (1931), pp. 308–434.

that have been raised not only by Buddington but also by H. L. Alling[3] and W. J. Miller.[4] In the light of new information, some of the writer's conclusions of 1931 must be corrected, others will probably have to be modified; but with regard to several fundamental questions he feels that no change of interpretation is as yet warranted. However, a detailed discussion of these must be postponed until additional field work has been completed.

AGES OF ADIRONDACK GABBROS

Observations on many gabbro bodies in the eastern Adirondacks led the writer to the conclusion[5] that these rocks are not younger than the syenite series, as had been advocated by several geologists, but that the large, massive bodies are older than the syenite and of approximately the same age as the anorthosite. In addition, a number of observations were presented which to him suggested a series of progressive evolutionary stages in the development of gabbros during the process of magmatic differentiation. It was thought that relatively small, layer- or lens-shaped gabbro masses represented initial stages in a clustering or compacting process which was believed to lead, through expulsion of interstitial, more siliceous materials, to larger and more basic gabbros. Although stocks, pipes, plugs, and laccoliths of gabbro had been mentioned in the literature, persistent search for such younger, intrusive gabbros was unsuccessful. In virtually all instances the contact relations were either contrary to what was to be expected or inconclusive.

The writer is therefore under great obligation to Dr. Buddington for having called attention to unquestionably discordant dikes of gabbro in anorthosite in the vicinity of Trembleau Mountain (Plattsburg-Willsboro quadrangles, New York) and in other localities farther south.[6] In 1942 the writer had an opportunity of visiting some of these exposures, and he fully confirms Buddington's observations. As the exposures, in the writer's opinion, are of great significance in Adirondack geology, he has mapped some of the dikes; and Figure 1 shows the outcrops in somewhat more detail than they appear on Buddington's map.[7]

The significance of these dikes is obvious. In at least one section of the anorthosite massif, truly gabbroic magma has been intruded into solidified anorthosite. The writer's conception that a series of evolutionary stages leads from gabbros *in statu nascendi* to large, spherical gabbros cannot be generalized for the eastern Adirondacks. Even if there have been large gabbro masses, segregating and settling out from a soft crystal mush, not yet crystallized throughout, there has been intrusion by seemingly identical magma. The writer may be criticized for ever having questioned this when many earlier investigators had claimed this relationship. However, after spending much time and effort in the field looking for precisely this evidence, he must say that the dike swarm of Trembleau Mountain is by far the most impressive and generally convinc-

[3] "The Adirondack Anorthosite and Its Problems," *Jour. Geol.*, Vol. XL (1932), pp. 193–237.

[4] "Emplacement of Adirondack Anorthosite," *Amer. Geophys. Union, Trans. of 1943* (1943), Part I, pp. 257–65.

[5] P. 341 of ftn. 2 (1931).

[6] Pp. 222–24 of ftn. 1 (1939); "Geology of the Willsboro Quadrangle, New York," *N.Y. State Mus. Bull. 325* (1941), pp. 49, 50, 53.

[7] Fig. 18, p. 223 of ftn. 1 (1939).

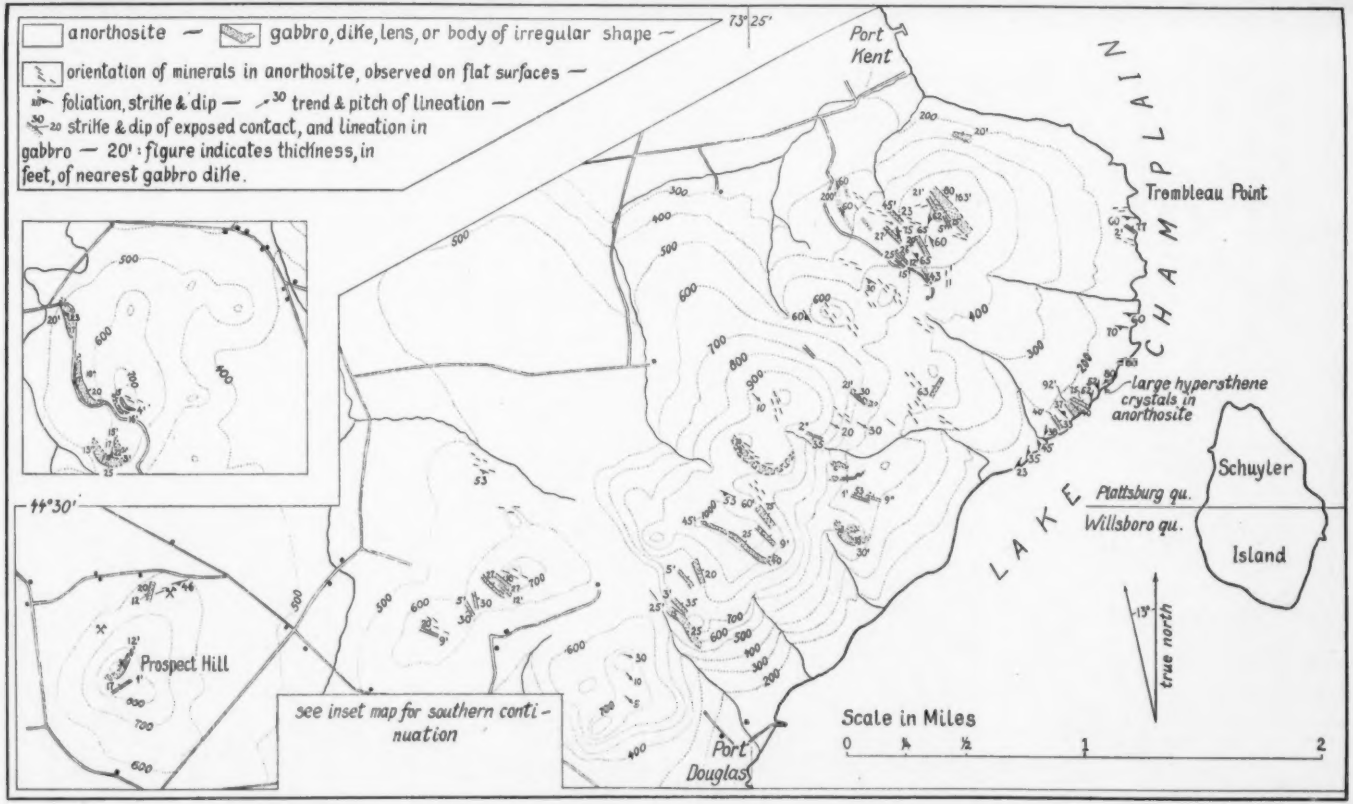
anorthosite — gabbro, dike, lens, or body of irregular shape —
orientation of minerals in anorthosite, observed on flat surfaces —
foliation, strike & dip — trend & pitch of lineation —
strike & dip of exposed contact, and lineation in gabbro — 20': figure indicates thickness, in feet, of nearest gabbro dike.
73° 25'
44° 30'
Port Kent
Trembleau Point
CHAMPLAIN
LAKE
large hypersthene crystals in anorthosite
Schuyler Island
Plattsburg qu.
Willsboro qu.
Prospect Hill
see inset map for southern continuation
Port Douglas
true north
Scale in Miles
0 ¼ ½ 1 2

FIG. 1.—Outline map of the vicinity of Trembleau Mountain, New York, showing location of discordant and concordant gabbro dikes in anorthosite. Dotted lines are contour lines as shown on the topographic quadrangle maps Plattsburg and Willsboro, New York.

ing case to prove the age relations. The possibility of other age relations of gabbros has always been kept in mind in the field; the single example of a foliated gabbro, cut discordantly by anorthosite, was published.[8] Specimens were also collected from a discordant medium- to coarse-grained dike, ½ mile south of Severance Hill, Schroon Lake quadrangle. In the field the rock resembled certain gabbros; but thin sections quickly proved this to be an exceptionally coarse diabase, as had been stated by Miller in the quadrangle report.[9]

There is, then, general agreement that anorthosite becomes locally gabbroic.[10] Balk's assumption[11] that from such rocks a series of rock types leads to the medium-grained, massive, commonly spherical gabbros (Buddington's "undersaturated gabbros"[12]) must be reinvestigated. If the latter are a "discrete group with no gradational relationships to the anorthositic rocks,"[13] it would, of course, be desirable to trace these two groups of gabbros throughout the area. It will prove an exceedingly difficult investigation. Even in the exposures at Trembleau Mountain, especially along the lake shore, where gabbro dikes are shown on Figure 1, there are beautiful "gradational" zones between gabbro and anorthosite. The upper contact of the 92-foot dike is fairly sharp and concordant with the anorthosite foliation; but below is a zone, 110 feet thick, in which anorthosite appears to grade into gabbroic films and thin layers—yet the evidence is admittedly strong that this sheet, like the others higher up on the mountain, is a true dike that was intruded into completely solidified anorthosite. Three thin sections were examined from this sheet. The upper and lower portions are fine-grained, well-foliated rocks; and 41 feet below the top is a layer, 13 feet thick, of medium-grained gabbro, with typical green, "dusty" plagioclases and garnet coronas. It lacks olivine, but this mineral is not present in all the dikes. The proportions of augitic pyroxene and hypersthene, as well as the distribution of hornblende and red biotite, are in every respect similar to corresponding features of the discordant dikes. On the other hand, it is almost unthinkable that a basalt or diabase dike would split near its contact into virtually paper-thin films throughout a stratigraphic thickness of over 100 feet. Some such process would probably have to be assumed to produce the transition zone after metamorphism of the rocks.

Similarly, we now have the question as to how some of the large spherical gabbros acquired their intensely schistose and lineated floors. These are a reality. The writer has demonstrated some of them to other geologists, on the occasion of the Sixteenth International Geological Congress and on other occasions. If it be assumed that the spherical gabbros were intruded, perhaps as laccolithic masses, into completely solidified anorthosite, and if they exerted such a force of expansion as to open up chambers as large as a cubic mile in some cases, one might expect a radial orientation of the lineation, away from a hypothetical central feeder. Field work in previous years did not show such an arrangement. In

[8] Fig. 29, p. 389 of ftn. 2 (1931).

[9] "Geology of the Schroon Lake Quadrangle," *N.Y. State Mus. Bulls. 213, 214* (1919), map.

[10] Buddington's "saturated gabbro group," pp. 214–16 of ftn. 1 (1939); Alling, p. 213 of ftn. 3 (1932); Miller, p. 258 of ftn. 4 (1943); "Significance of Newly Found Adirondack Anorthosite," *Amer. Jour. Sci.*, Vol. XVIII (1929), pp. 393–94.

[11] P. 405 of ftn. 2 (1931).

[12] P. 215 of ftn. 1 (1939).

[13] *Ibid.*, p. 215.

view of the importance of this problem, the writer intends to re-examine several spherical gabbros as soon as opportunity offers.

GABBROIC DIFFERENTIATES OR SKARN?

Buddington states[14] that some of the rocks called by the writer "gabbro *in statu nascendi*,"[15] may in reality represent Grenville skarn. It is quite possible that many dark-colored layers and zones in relatively dark anorthosite may, after microscopic examination, prove to be derived from Grenville lime-silicate rocks. Not every outcrop can be sampled, and not every hand specimen can be sectioned. Buddington himself says[16] that mafic anorthosite or, as the case may be, relatively mafic syenite may result from assimilation of Grenville rocks. In order to check on the probability of erroneous field identifications of rocks, the extensive railroad cuts north of Willsboro Station were re-examined. To judge from the results in hand, the danger of confusion is remote. Field identifications of gabbroic and lime-silicate rocks were amply confirmed by thin sections and powder identifications of minerals. The association of a diopsidic pyroxene with scapolite, in the latter group, and of an augitic pyroxene with hypersthene, in the former, seems diagnostic in most specimens. Of course, amphibolitic rocks are more troublesome. However, very few of the thin gabbroic layers develop this mineral in excess, although this may be different in other sections of the anorthosite massif.

FIG. 2.—Flat lens of fine-grained gabbro, concordantly folded in Grenville marble and lime-silicate rock. Southern entrance to tunnel, Delaware & Hudson Railroad track, north of Willsboro, New York.

Five localities were sampled:

1. Southern tunnel entrance, on the Delaware and Hudson Railroad track, about 3 miles north of Willsboro Station (Willsboro quadrangle, New York). About 70 feet from the tunnel entrance, on the west side of the track, is a flat lens

[14] Pp. 44–46 of ftn. 1 (1939).

[15] P. 351 of ftn. 2 (1931).

[16] *Ibid.*, pp. 43, 47, 128; Buddington, pp. 34, 37 of ftn. 6 (1941).

of fine-grained gabbro (Figs. 2, 3). It is conformably folded into Grenville marble, and a narrow horizontal wedge of the latter in the northern front of the gabbro is probably the remnant of a tight, recumbent fold. A few inches above the upper gabbro contact, a layer of fine-grained reddish-gray rock is seen, which was considered a lime-silicate rock. Seven specimens and thin sections are located as follows (Fig. 3): (1) upper gabbro contact where lime-silicate rock almost touches it; (2) lime-silicate layer, 2 inches above its base; (3) gabbro, 10 inches below upper contact; (4) gabbro, 3 inches below upper contact; (5) gabbro, 18 inches below upper contact; (6) gabbro, ½ inch below upper contact; and (7) gabbro contact. In its interior the lime-silicate lens is a fine-grained aggregate of closely interlocked garnet and quartz. The garnets are of ragged, spongiform shape; and quartz occurs both as minute blebs in the groundmass and as large single grains, scattered through the matrix. Associated with garnet and quartz is a fine-grained micaceous aggregate, in which an occasional remnant of scapolite suggests the source. Diopside is largely altered, but a few remnants are identifiable. Carbonate occurs as sporadic grains. The upper contact of the gabbro is sharp (Fig. 4). The gabbro ends abruptly, as a straight "front" of brown hornblende and magnetite. Above the gabbro contact there is a zone, about 1 inch thick, composed of light-green pyroxene, scapolite, sphene, and calcite. Sphene is particularly abundant within 5–7 mm. above the gabbro. Beyond the 1-inch zone the amount of diopside decreases rapidly, and scapolite also is less abundant. This sudden drop is responsible for a conspicuous change in color. Megascopically, the diopside selvage might be mistaken for gabbro, although even in the hand specimen the coarser grain and greenish-gray color distinguish it from the darker and finer-grained gabbro. In section No. 7 the gabbro minerals at the contact include plagioclase and apatite, and some grains

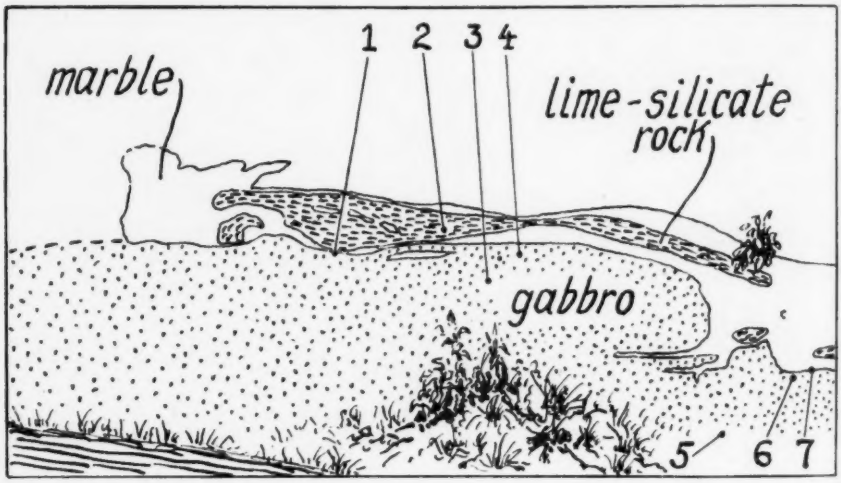

FIG. 3.—Same exposure as shown in Fig. 2, indicating location of specimens collected

of scapolite extend about 2 mm. into the gabbro. Lenses of andesine, as well as some large grains of microperthite, were also noted. The gabbro is composed of andesine-labradorite (ab_{50}), augitic pyroxene, smaller amounts of hypersthene, and variable proportions of brown hornblende, biotite, garnet, ore minerals, apatite, and potash feldspar. Apart from the tendency of scapolite to enter the gabbro border in small amount, the mineral associations of lime-silicate rocks and gabbro appear to be distinct.

2. Steep north-facing cliff, 450 feet south of the southern outlet of the same tunnel.[17] Six feet above what the writer had called a gabbro *in statu nascendi* is a row of dark lenses which differ megascopically from the gabbroic layer through their deep-green, almost black, color. In the photograph referred to in footnote 17 they appear at the upper right-hand corner and farther along to the left. The caption of this photograph says "gabbroic layers," because it was considered that not only the bottom layer in the cliff but also the numerous dark films and thin layers (but not the lenses just referred to) represented material in the process of settling and gathering. Six thin sections were prepared from a vertical zone, about a foot thick, which included the lime-silicate lens and the layered anorthosite above and below. One thin section was cut across the top of the gabbro *in statu nascendi*, which lies 5 feet 4 inches below the lime-silicate lens. Unfortunately, the anorthosite with thin dark layers proved extensively altered. Sericite and prehnite have all but destroyed the feldspar, and original red biotite has largely been chloritized. However, the mafic minerals which constitute the thin dark films are better preserved. They consist of a pale-green pyroxene, garnet, green hornblende, and some magnetite. Apatite and quartz are also conspicuous in some layers. The lime-silicate lens is a medium- to fine-grained, dark greenish rock in which scattered poikiloblasts can be seen, about ½ inch across. They are probably scapolite, but the mineral could not be isolated for identification. The mineral composition in thin section is: almost colorless pyroxene (α, 1.693; β, 1.702; γ, 1.722), about 40 per cent; alteration products of scapolite, 30 per cent; hornblende, about 20 per cent; sphene, 5 per cent; calcite, a little plagioclase, garnet, and quartz, 5 per cent. Although sphene and scapolite (or its alteration products) could not be identified in the thin dark films that surround the lens, there is the possibility

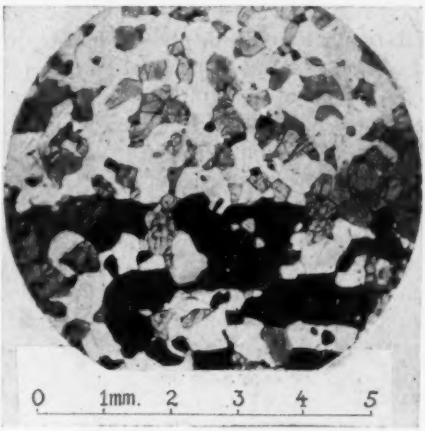

FIG. 4.—Photomicrograph of upper contact of gabbro with Grenville sediments, as shown in Figs. 2 and 3. Contact of gabbro (lower half) marked by brown hornblende (large dark grains) and magnetite (small black grains), with smaller amounts of slightly pleochroic green pyroxene (gray) and plagioclase (white). Lime-silicate rock above gabbro consists of light-green pyroxene (large gray grains), sphene (small gray grains), scapolite (white), and a little magnetite (black). Plain polarized light.

[17] Balk, Pl. 5, photo 14, p. 348, of ftn. 2 (1931); Buddington, p. 45 of ftn. 1 (1939).

that the pale-green pyroxene is due to incorporation of sedimentary material in the anorthosite. However, the mineral association in the gabbro *in statu nascendi* differs appreciably from the overlying rocks. The pyroxene is deeper green ($\beta = 1.704$), with well-developed diallage structure. The plagioclase is entirely unaltered and contains occasionally antiperthitic blebs of potash feldspar. Greenish-brown hornblende and red biotite, in addition to about 35 per cent garnet, are the principal mafic minerals. Hypersthene is reported by Buddington; in the thin section from the gabbro top it is lacking, but its presence is in accordance with the usual composition of gabbros.

3. Milepost A144 RP47 is located about 300 feet north of the railroad maintenance cabin, where the lake shore first approaches closely the railroad, north of Willsboro Station. In anorthosite directly opposite the milepost, on the west side of the track and about $5\frac{1}{2}$ feet above it, there are three lens-shaped masses of dark rock, 3–5 inches across. The southern two, considered to be lime-silicate rock, have a grayish interior and a darker green fringe. The northernmost lens appears to be more mixed up with anorthosite, and amphibole seems the only mafic mineral. Only small fragments could be secured here, since the vertical ledge surface is quite smooth. Nevertheless, the powdered dark grains from the interior of the southern two lenses have the properties of diopside, whereas hornblende is the only mafic mineral from the fringe and from the northern lens.

4. For about 1,000 feet north of milepost A144 RP47 the railroad track is straight. At the north end of this straight stretch the base of a large, flat, fine-grained gabbro sheet is exposed on the west side of the track. It is floored by anorthosite, with many andesine augen, $\frac{1}{2}$–1 inch across, having much mafic material between them. Nothing suggests Grenville silicate rock. Four thin sections were cut from the floor zone: one from the gabbroic rock 12 inches above the floor; one through the gabbro floor; one through dark anorthosite films, $2\frac{1}{2}$ inches below the gabbro floor; and the fourth through the normal mafic anorthosite, 4 inches below the floor. The gabbro contact, as seen in the thin section, is sharp. Continuous layers of augitic pyroxene and ore abut the anorthosite; only garnet occurs uniformly in both rocks. The gabbro contains about 25 per cent augite; 15 per cent hypersthene, plus secondary iron ore derived from its alteration; about 15 per cent garnet; and about 40 per cent andesine (ab_{52}), with small amounts of potash feldspar, red biotite, hornblende, and apatite. Light-green augite, garnet, and a trace of altered hypersthene are the mafic minerals in the anorthosite; hornblende was seen only in the darker film. Potash feldspar in the anorthosite ranges up to 12–15 per cent; calcic andesine (ab_{52}) amounts to about 35 per cent.

5. About 75 feet north of locality (4), where the track makes a curve to the northwest, two pinkish-green lenses of lime-silicate rock are exposed at a height of about 7 feet above the track. They are surrounded by anorthosite with many lenticular pegmatite pods, and above them follows a zone of dark gabbroic anorthosite. Five thin sections were prepared through the different rock types. In the lenses of lime-silicate rock, diopside, sphene, and scapolite are the chief minerals, whereas the dark gabbroic anorthosite above is composed of augite, hornblende, garnet, and chloritic aggregates, probably altered hypersthene. As is often the case, potash feldspar and

microperthite are conspicuous in the boundary zone and amount to about 9 per cent of the rock, in addition to about 20 per cent of andesine (ab_{57}).

Buddington[18] refers to two other localities where rocks called gabbro *in statu nascendi* by Balk are considered to be skarn rocks. One is a small road cut on highway No. 22, exactly 1 mile due east of the letter *L* in the word "*L*ong Pond," Willsboro quadrangle, New York. The specimen which was collected comes from the south side of the highway, at the level of the roadbed, exactly opposite the point where a footpath goes off to the farm on the north side of the highway. The flat-lying layers of anorthosite here contain several dark-gray to reddish-gray layers, schlieren, and ill-defined concordant zones, in which red garnet and black biotite are identified with the naked eye. Andesine is abundant in all layers; and the dark zones contain, in addition to garnet and biotite, some brown hornblende. Pyroxene is lacking in the specimen examined. It is admitted that this lack of pyroxene may speak against inclusion of this layer with gabbroic rocks (it was called by Balk a "highly garnetiferous gabbro amphibolite");[19] however, Buddington's description leaves no doubt that he has examined an entirely different rock.

The second locality is a quarry, $2\frac{1}{2}$ miles south of Schroon River (Paradox quadrangle, New York). The rock was called a "garnet-bearing gabbroamphibolitic flow layer in anorthosite,"[20] and a chemical analysis was given. Buddington writes:

> The writer's study of a specimen from this locality shows it to consist of a granoblastic aggregate of andesine (ab_{57} an_{43}), brownish hornblende (γ: 1.689, α: 1.669), and garnet with accessory augite, hypersthene, and magnetite. The outcrop shows no independent evidence that this is a layer of Grenville gneiss formed by metamorphism of limestone and subsequently recrystallized, but its characters are consistent with such an interpretation.[21]

The present writer is unable to understand what characteristics of this rock, composed as it is of the normal minerals of the anorthosite and without any critical lime-silicate minerals, are to be interpreted as skarn. Hypersthene is, in our experience, extremely rare in Grenville lime-silicate rocks, whereas it is one of the standard minerals of gabbros and gabbroic anorthosite. The composition of the plagioclase is likewise the normal one for anorthosite-gabbro.

On the basis of admittedly inadequate observations, the writer believes that the danger of confusing in the field Grenville lime-silicate rocks with gabbroic facies of the anorthosite is not great. On the other hand, the question as to how far the chemical and mineral composition of gabbroic anorthosite may have resulted from assimilation of sedimentary rocks deserves further careful study. Although, owing to Buddington's comprehensive work, there is now available a fair number of good chemical analyses of the principal rock types, we still lack analyses of the common rock-forming minerals, especially in the pyroxene, hornblende, and garnet groups. Detailed investigations of localities where Grenville lime-silicate zones pass into gabbroic anorthosite, preferably along the strike, would probably be of great value.

THE ROLE OF VOLATILES IN ANORTHOSITE

In 1931 the writer described exposures of anorthosite where mafic minerals

[18] P. 45 of ftn. 1 (1939).

[19] P. 350 of ftn. 2 (1931).

[20] *Ibid.*, p. 425.

[21] P. 45 of ftn. 1 (1939).

(usually pyroxene, with or without minor amounts of hornblende and biotite) appear clustered around the edges of blocks of anorthosite.[22] Occasionally, the andesine crystals inside such blocks are larger than those outside. At the time, this was considered as evidence that ferromagnesian minerals had been conveyed in a crystal mush and had been arrested along the edges of a more completely solidified block of anorthosite. Although this interpretation may still hold true in many instances, the writer has since satisfied himself that another interpretation must also be considered, namely, that the fringes of dark minerals, and possibly even some of the large andesines, may have developed from pneumatolytic processes and thus may not be directly referable to magmatic movements, although the latter may have indirectly facilitated the growth of larger crystals.

Such a possibility was first brought to the writer's attention in the Egersund anorthosite massif in southern Norway, which he had the pleasure of studying in the summer of 1937, in the company of Drs. Tom F. W. Barth and Anders Kvale. Most of the Egersund anorthosite is a medium-grained rock that resembles outwardly certain nepheline syenites. Plagioclase augen are rather rare; but here and there the eye is struck by immense pyroxene crystals, as much as a cubic foot in volume, lying in normal anorthosite, either singly or in groups of several crystals (Fig. 5). Occasionally the crystals lie in a straight row, without touching each other and without a visible crack between them; no other minerals seem to be associated with them.

FIG. 5.—Group of single crystals of pyroxene in anorthosite, northwest of Bjerkrem, Egersund massif, Norway.

The appearance of such large crystals is certainly suggestive of pneumatolytic processes, and the question as to why there are single crystals rather than regular dikes or lenses of other coarse minerals leads to a number of speculations. The writer was interested, therefore, to

[22] Figs. 15–17, pp. 353–55 of ftn. 2 (1931).

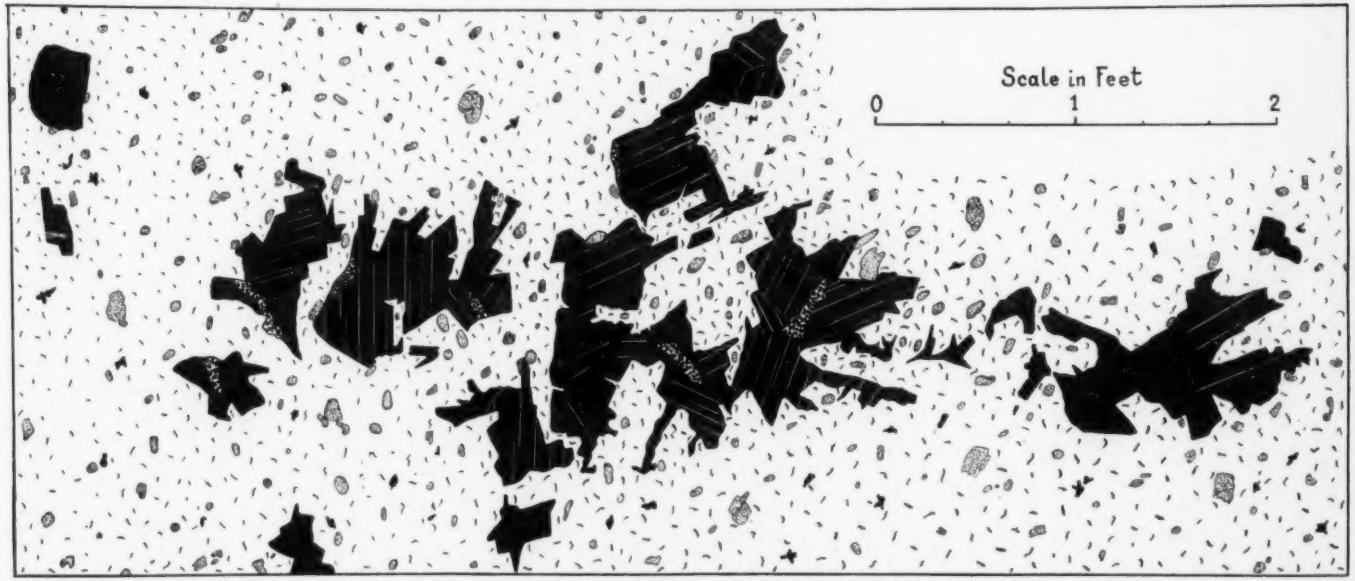

FIG. 6.—Large crystals of pyroxene, believed to have been precipitated from volatiles, in ledge of anorthosite, west shore of Lake Champlain, northwest of Schuyler Island (Plattsburg quadrangle, New York). Black: pyroxene, thin white parallel lines indicating size of individual crystals as established by cleavage reflections; coarse stippling inside pyroxene crystals: magnetite; fine stippling outside: andesine augen in anorthosite.

find similarly large crystals of pyroxene in anorthosite at the shore of Lake Champlain, northwest of Schuyler Island (southern portion named "Whitney Island" on the new 1943 edition of Willsboro quadrangle) (Figs. 6, 7). At a point in a direction N. 55° W.[23] from the northern end of the island (Fig. 1) the rather massive anorthosite, with andesine augen of moderate size, is peppered with large pyroxenes. The crystals have ragged terminations and resemble poikiloblasts to some extent. In the sunlight, crystals measuring over 6 inches in length give simultaneous cleavage reflections, with contiguous crystals separated from each other by an inch or more of anorthosite. In places the pyroxene boundaries follow andesine cleavages (not recognizable in Fig. 6); and interstitial seams or cross-cutting veinlets of magnetite (probably titaniferous) are numerous. At another point on the lake shore, in a direction N. 68° W. from the north end of Schuyler Island, there is a squarish mass of structureless anorthosite, somewhat more than a square foot in size, in which andesine crystals, 1–2 inches across, are associated with large pyroxenes of the same poikiloblastic habit, as noted above, and a little magnetite (Fig. 7). The surrounding anorthosite is slightly foliated and lacks the large plagioclases. About 80 feet south of this locality the anorthosite displays truly immense pyroxene crystals (locality indicated on Fig. 1). A surface measuring 18 × 25 feet is criss-crossed by several dozen crystals as large as 8 inches in width and 32 inches in length. The crystals are hypersthene, with the following properties: α, 1.684; β, 1.692; γ, 1.696. Pleochroism: X = wine red, Y = pinkish-yellow, Z = pale green; 2V large, optically negative, $\rho > v$, distinct, Z:c = 0°. According to N. H. and A. N. Winchell,[24] the composition is about 21 per cent enstatite, 79 per cent $FeSiO_3$. Some of the crystals are marginally replaced by chlorite, and irregular spots and seams of magnetite are common.

The location of this instructive exposure is as follows: Walk on the Delaware and Hudson Railroad track south from Port Kent (Plattsburg quadrangle) to milepost A153 RP38. Continue south to the second cut, about 1,100 feet from milepost, with prominent brick-red south-facing joint surfaces. About 150 feet south of this cut is a platform on the east side where maintenance crews leave their motor trucks. Walk southeast from platform to shore ledges—a distance of about 120 feet.

Farther south on the lake shore, where a brook enters from the northwest, large blocks have been dumped from a near-by railroad cut. Some of these, too, exhibit large crystal masses of a green augite (α, 1.689; β, 1.695; γ, 1.711) which are penetrated by many dull-green zones of fine-grained, recrystallized augite of the same composition and also by veinlets of calcic andesine (ab_{52}). Associated with these are irregularly shaped masses of magnetite, pyrrhotite, pyrite, and lavender-colored apatite.

In view of these observations, to which the writer hopes to add in the near future, it seems as though the effect of volatiles in the anorthosite massif is perhaps greater than might seem apparent at first. It is true that the role of volatiles in the anorthosite has long been recognized, and pegmatites in the rock have often been described.[25] On the other

[23] Compass directions quoted in this paper are not adjusted for declination.

[24] *Elements of Optical Mineralogy* (2d ed.; New York: John Wiley & Sons, Inc., 1927), p. 177.

[25] Alling, pp. 203–6 of ftn. 3 (1932); Buddington, p. 35 of ftn. 1 (1939) and p. 48 of ftn. 6 (1941); J. F. Kemp and R. Ruedemann, "Geology of the Eliza-

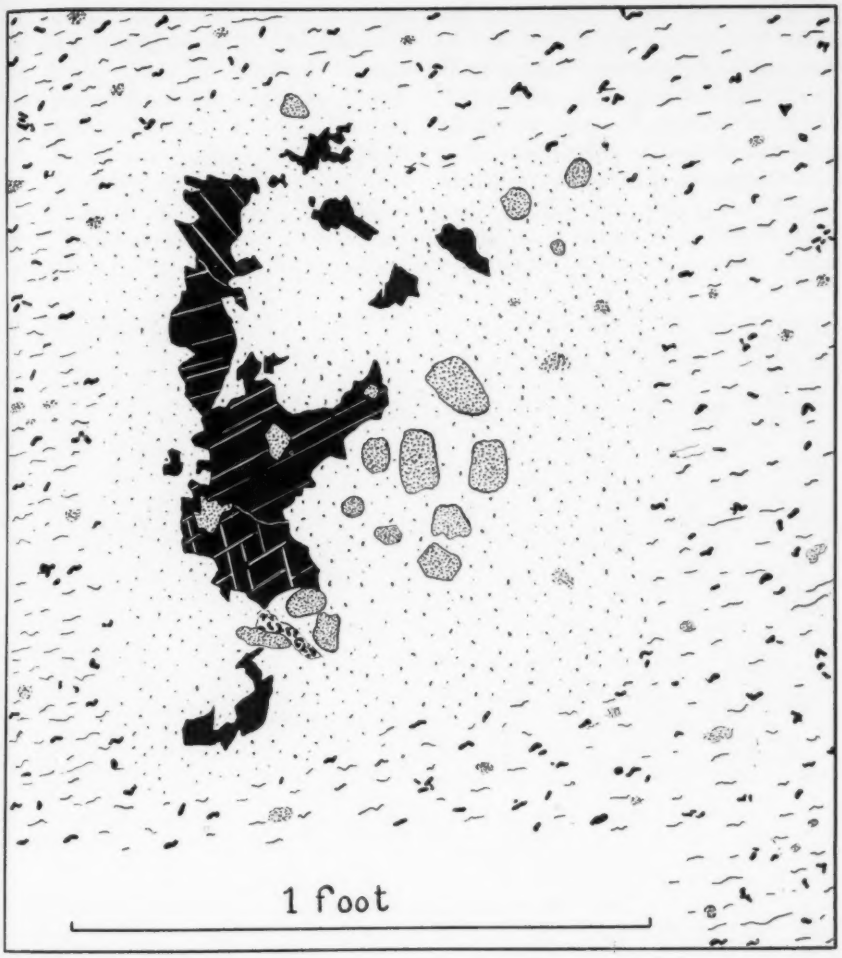

FIG. 7.—Mass of compact anorthosite with large, ragged pyroxene crystals and several large andesine augen, surrounded by slightly foliated anorthosite. Note small veinlet of magnetite in lower left portion (coarse stippling). Lake Champlain shore, northwest of Schuyler Island.

hand, a pegmatite exposure is conspicuous, whereas isolated or weathered pyroxene crystals, a few inches across, may escape notice in till-covered and forested country. After seeing the outcrops at the Lake Champlain shore, the writer paid more attention to large pyroxene crystals in the anorthosite, and several localities were noted. For instance, at the extreme northern base of the northeast

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bethtown and Port Henry Quadrangles," *N.Y. State Mus. Bull. 138* (1910), p. 30; I. H. Ogilvie, "Geology of the Paradox Quadrangle, New York," *N.Y. State Mus. Bull. 96* (1905), p. 488; Kemp and Alling, "Geology of the Ausable Quadrangle," *N.Y. State Mus. Bull. 261* (1925), p. 39.

spur of Mount Bigelow (Willsboro quadrangle, New York), where an old road skirts the mountain, hypersthene crystals up to a cubic inch in size are not rare. A row of pyroxene crystals traversing coarse anorthosite is well exposed in a blasted ledge on the north side of the highway from Upper Jay to Wilmington (Lake Placid quadrangle, New York), about 2 miles northwest of Upper Jay (Fig. 8). Pyroxene crystals attain about 1 cubic inch in size, and many smaller specks of dark minerals surround them as an ill-defined outer zone. The entire zone, including several inches of the surrounding anorthosite, has been intensely saussuritized and chloritized, so that the exact composition of the original minerals cannot be verified, except for the cores of the largest pyroxenes. The bluish andesine augen of the anorthosite become chalk-white toward the edges of the zone, and in its center some pyrite has crystallized. Pyroxene crystals that have followed the border of a fine-grained, whitish anorthosite block are shown in Figure 9. The crystals that have grown along the fractures are small but extend with angular terminations into the surrounding anorthosite. Inside the block pyroxene crystals attain 2 inches in length; but, on the whole, the right block is poor in this mineral, whereas the outer anorthosite contains more mafic minerals than is normal. This raises the important question of whether all the mafic minerals in the left portion of the outcrop were precipitated from volatiles or whether there are two generations of them: an early one, crystallized with the surrounding andesine during the cooling of the rock, whatever the exact process may have been; and a later generation precipitated from volatiles either in pore spaces or in cracks produced perhaps by crushing of the rock.

FIG. 8.—Row of pyroxene crystals in massive, coarse-grained anorthosite. Highway to Wilmington, 2 miles northwest of Upper Jay, Lake Placid quadrangle, New York. The surrounding anorthosite is saussuritized, and the margins of pyroxene crystals are chloritized.

It is worthy of note that andesine has also been precipitated in places from the

same hyperfusibles that precipitated pyroxene. Andesine veinlets in augite at the Lake Champlain shore were mentioned above. They crosscut the pyroxenes which themselves would seem to have replaced the anorthosite. Laths of labradorite (ab_{48}), as much as an inch in length, penetrate augite crystals (α, 1.687; β, 1.692; γ, 1.713; 2V about 65°; optically positive, $\rho > \upsilon$, weak) in a remarkable gabbroic pegmatite on the high crest west of summit 3601′, Jay Mountain (Ausable quadrangle, New York) (Fig. 10). The coarse rock consists of augite crystals, large cleavage fragments of which strew the mountainside. In the ledges along the ridge crest large poikiloblasts of pyroxene enclose many large plagioclase crystals. There seem to be two generations here: an older one, of bluish color, occurs in large crystals up to matchbox size, whereas a younger generation forms small, yellowish-white grains and zones. The pyroxene crystals seem to be older than the second plagioclase generation but younger than the large bluish ones. Along the edges of the latter the enclosing pyroxenes develop thin selvages of biotite, garnet, hornblende, or any combination of these, as well as magnetite.

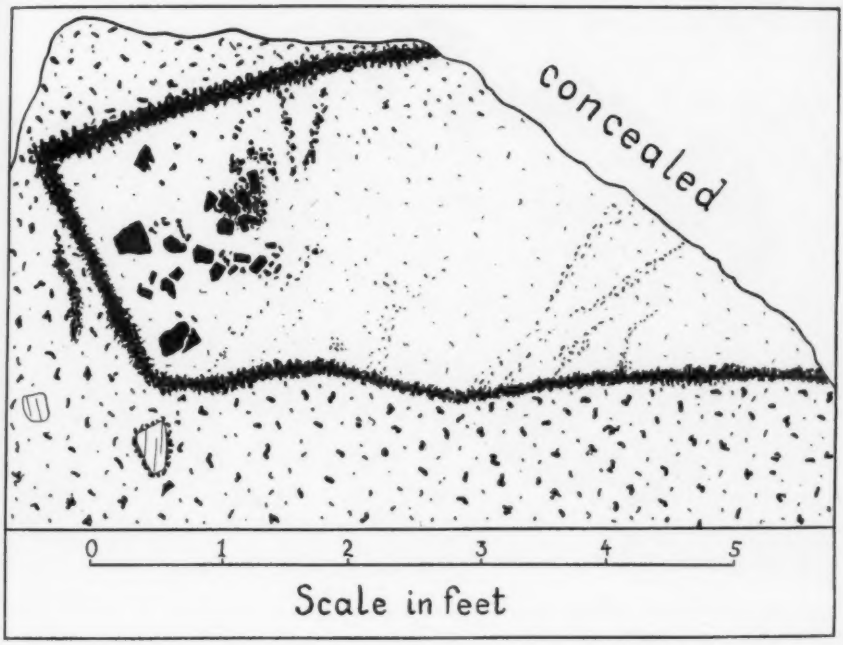

FIG. 9.—Glaciated surface, 180 feet east of summit 560′, Trembleau Mountain, Plattsburg quadrangle, New York, showing large pyroxene crystals in a mass of whitish, fine-grained anorthosite. Marginal rows of smaller pyroxenes exhibit crystal terminations as though pyroxenes have replaced anorthosite. The surrounding anorthosite contains more pyroxene than the central mass. Note pyroxene rim around andesine auge in lower left corner.

If the writer's scattered observations are correct, the well-known protoclastic granulation of the anorthosite is older

than the precipitation of the large pyroxenes, but much more field work is required to establish this.

These few observations may not throw much light on the problems here under discussion; yet there is little doubt that at least in some places mafic minerals in the anorthosite have developed to impressive size through volatiles. The composition of these volatiles remains a matter of speculation, although iron oxide, phosphoric acid, and probably titania must have been present, in addition to silica and other common components. Important is the question as to what may have been the volatile compound that precipitated titania. In the Egersund anorthosite massif there are straight, crosscutting ilmenite dikes (Fig. 11) and others composed of ilmenite and coarsely crystalline hypersthene and andesine-labradorite. For the nelsonite

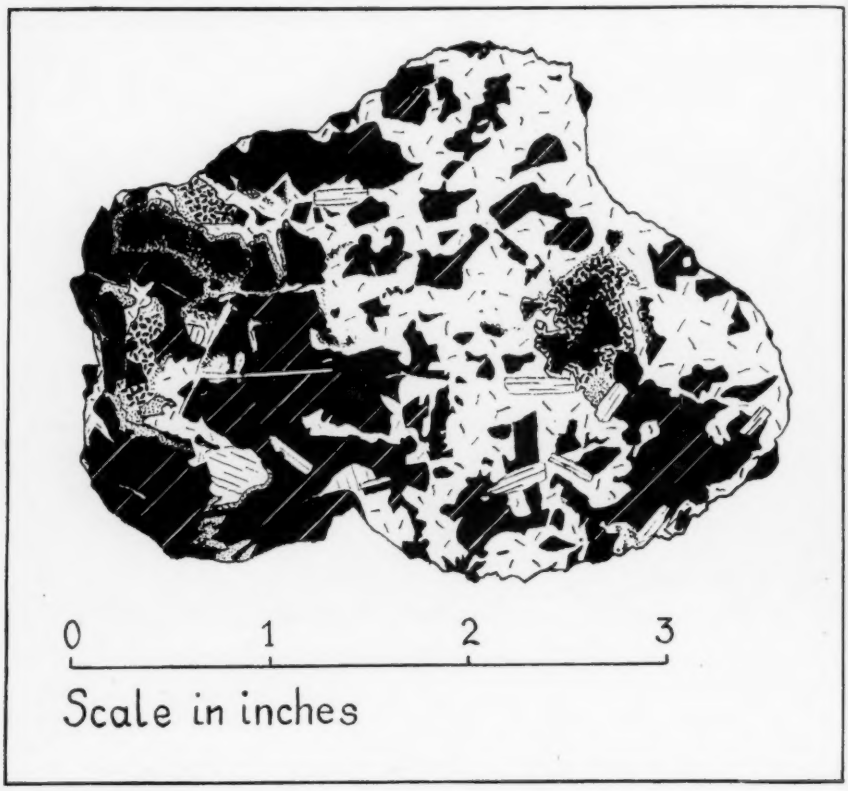

FIG. 10.—Drawing of hand specimen of gabbroic pegmatite from crest of Jay Mountain, west of summit 3601′, Ausable quadrangle, New York, showing poikiloblastic growth of augite crystals (black). Two thin, tabular crystals of plagioclase, in lower left corner, seem to replace the augite. Parallel, thin white lines in pyroxene indicate identical orientation of cleavages. Coarse stippling: hornblende and biotite; fine stippling: garnet. Several large, bluish andesine augen are shown by parallel lines.

dikes in Virginia, C. S. Ross[26] assumes titanium tetrafluoride because of the presence of fluorine in small amounts in some of the surrounding minerals. Detailed mineralogical studies in the Adirondacks are much needed to clarify the nature of these mineral associations and to provide quantitative data from which the course of differentiation and fractionation of the magma can be re-

and others[28] one would infer that the anorthosite developed fractures locally. They gaped long enough to permit the precipitation of minerals along them but seem to have closed rather soon, so that minerals of only one species were precipitated in some of them. It may even be that the walls contained enough of a pore liquid to allow a seamless freezing-together of the closing walls. This might

FIG. 11.—Five-inch ilmenite dike in anorthosite. North Sea coast at Kalvshagen, Egersund anorthosite massif, Norway. The straight, crosscutting dike, near feet of man, is commercial ore. Chunks of ore are chipped out with wedges and sold by local farmers.

constructed. An interesting suggestion along this line has been made recently by Dr. Buddington.[27]

Structurally, the question arises as to the mechanical condition in which the anorthosite may have been when the replacement occurred. From Figures 8, 9, give us the straight rows of pyroxene crystals (p. 298). The process pictured here agrees closely with Ross's conception of the emplacement of the dike-like ore bodies in the Roseland district, Virginia.[29]

Observations on other types of veins or dikes throw additional light on this stage in the crystallization of the anorthosite. Very common are dike-like masses, composed of garnet, microperthite, andesine, and quartz, with variable amounts

[26] "Occurrence and Origin of the Titanium Deposits of Nelson and Amherst Counties, Virginia," *U.S. Geol. Sur., Prof. Paper No. 198* (1941), pp. 35–36.

[27] "A Course of Concentration of Iron, Titanium, and Phosphorus in Magmatic Differentiation" (abstr.), *Amer. Geophys. Union, Trans. of 1942*, Part II (1942), p. 346.

[28] Balk, Figs. 4, 5, and 15–17 of ftn. 2 (1931).

[29] P. 18 of ftn. 26 (1941).

of potash feldspar, hornblende, biotite, and pyroxene. In contrast with aplite or diabase dikes, these rarely appear to extend over long distances; their widths are variable, a vein a foot thick here may dwindle to a few millimeters within 30 feet; some of them cross massive anorthosite; others are concordant in foliated rock; still others are discordant part of the way. The fact that many of these veins are intensely schistose, whereas the surrounding anorthosite may be but faintly foliated, or massive, was emphasized in 1931.[30] The earlier observations were amply confirmed in 1942, and many new veins were sampled and sectioned. Particularly interesting is the tendency of some of them to come to vague, flaring ends, somewhat like a cornucopia (Fig. 12). In 1931[31] it was emphasized that a relatively siliceous fluid must have moved along these channels, and the hypothesis was entertained that this may have been a portion of the mother-liquor. At any rate, the exposures leave little doubt about a period in the consolidation of the anorthosite when rupturing and flowage overlapped. This picture seems to call for amplification in the light of the outcrops described above. It appears that, in addition to expulsion of a liquid fraction, there has been also some permeation of the congealing anorthosite by volatiles, precipitating in places the normal minerals of the anorthosite, occasionally in crystals of exceptional size.

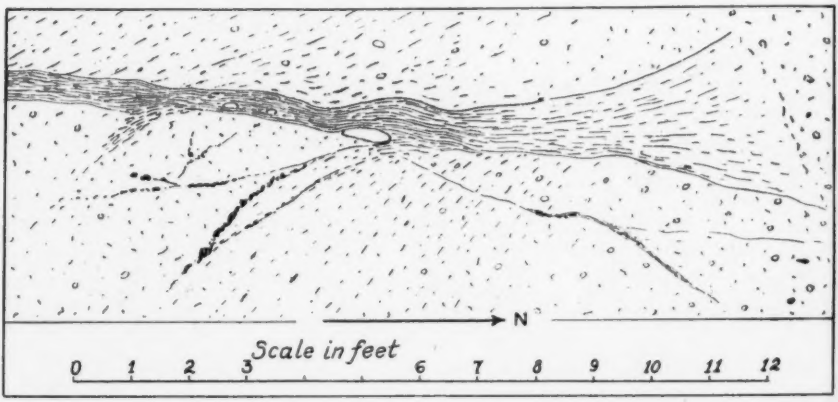

FIG. 12.—Siliceous dike, partly discordant, in slightly foliated anorthosite. Glaciated ledge, east-southeast base of Prospect Hill, Willsboro quadrangle, New York. The dike is composed of 40 per cent microcline, 30 per cent quartz, 18 per cent andesine (ab_{52}), 10 per cent garnet, and 2 per cent accessories (brown biotite, green hornblende, apatite, zircon). Note flexure in bordering anorthosite suggesting that the rock was still soft and sheared by flowage when the dike was emplaced. On the right the dike widens to 47 inches, ending blindly in anorthosite. Dike contacts dip at 75° (west side) and 45° (east side) to the west.

ANORTHOSITE DIKES

The postulated stage of overlapping fracturing and flowage in the presence of volatiles may help to explain the peculiar nature of the much-discussed anorthosite dikes. Although there are scattered references in the literature to anorthosite dikes, they seem to be remarkably few. In the Adirondacks the writer has yet to see a dike of any appreciable

[30] Pp. 361–62 of ftn. 2 (1931).

[31] *Ibid.*, pp. 369–72.

length and one with straight contacts. However, there are some perfect dikes in the Egersund massif in Norway. The dike reproduced in Figure 13 is as straight as any granite dike can be; its contacts are sharp; they cut across faintly foliated anorthosite at an angle of 75°; the dike can be walked out for over 200 feet. The anorthosite inside the dike is feebly foliated parallel to the contacts, but the dike rock and the surrounding anorthosite are indistinguishable in specimens. Were it not for the discordance of the two foliations, one might step across the contact and not know there was a dike. In the Egersund massif, which is as perfectly exposed as the coast of Greenland or Labrador, we saw four anorthosite dikes. Some have more mafic minerals than the anorthosite; others not. Although many miles of the contact of the massif were walked out, not a single apophysis of the anorthosite was seen. Foliated anorthosite in the Egersund massif becomes massive toward the contact. The surrounding gneisses are tilted and disturbed; but nowhere can the age relations with the anorthosite be established, although the contacts we saw are apparently not fault contacts. It seems, therefore, as though this rock is capable of intruding itself but not the wall rocks. This anomalous behavior might be somewhat easier to understand if it is assumed that the cooling proceeded very slowly and that, as a consequence, there was an overlapping in time of incipient fracturing and a diminishing amount of flowage. This condition might permit the occasional filling of a temporary fissure with anorthosite mush, either of the same texture as the immediately surrounding rock or, if volatiles were present, with slightly higher con-

FIG. 13.—Discordant dike of medium-grained anorthosite, 5 feet 4 inches wide, cutting medium-grained, slightly foliated anorthosite with sharp, straight contacts. Arrows indicate dike contacts. North Sea coast, ¼ mile southeast of Odden, Egersund anorthosite massif, Norway.

tents in mafic minerals than the surrounding rock. Where volatiles precipitated mafic minerals in an early formed fracture, while the surrounding anorthosite was still capable of flowage, the veneered fractures could be sheared out, with accompanying flexure zones in the surrounding anorthosite.

Many examples could be cited that would fit such a development of the

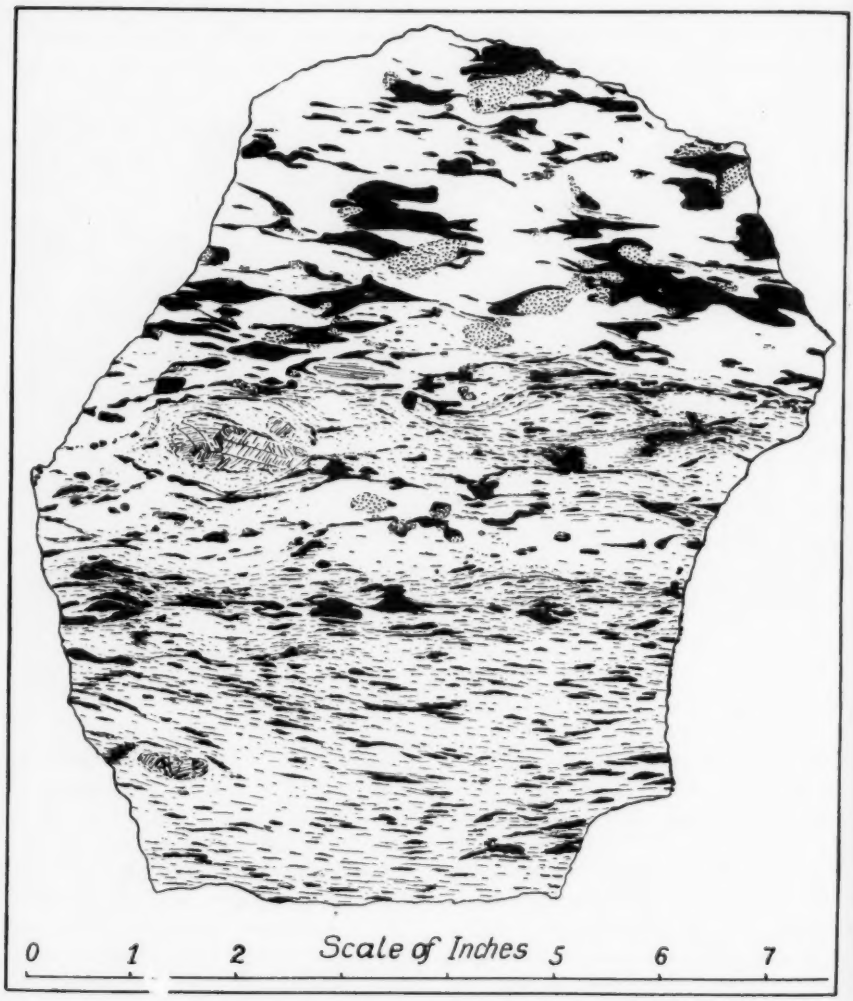

FIG. 14.—Tracing of specimen of foliated augite-hedenbergite-sphene zone in anorthosite. Black: hedenbergite in upper portion, augitic pyroxene in lower layers; stippling: sphene; white: feldspar. Two andesine augen in left portion of specimen. Abandoned road metal quarry on highway 9N, 2 miles northeast of Upper Jay, Lake Placid quadrangle, New York.

veins. One may be briefly referred to. Buddington[32] describes a sphene-gabbro pegmatite from the south shoulder of Hamlin Mountain (Lake Placid quadrangle, New York). The dike is composed of andesine, augite, and sphene, plus accessory minerals. The minerals are reported to be oriented with their longest axes perpendicular to the dike wall; in other words, no later deformation has crushed the original texture of the dike. However, $2\frac{3}{4}$ miles due south of this locality, in an abandoned road-metal quarry on the northwest side of highway 9N, 2 miles northeast of Upper Jay (eastern border of Lake Placid quadrangle, New York), a very similar mineral association, with a noteworthy concentration of sphene, is found in a strongly foliated zone, $4\frac{1}{2}$ feet thick, which passes upward and downward into apparently normal anorthosite (Fig. 14). The pyroxene in the upper layers is light-green augite; but in the lower portion, which is coarser grained and carries most of the sphene, the pyroxene is strongly pleochroic hedenbergite (α, 1.722; β, 1.728; γ, 1.743; i.e., about 77 per cent hedenbergite and 23 per cent diopside, according to Winchell);[33] the other minerals include calcic andesine (ab_{53}), commonly with antiperthitic blebs of potash feldspar, sphene, apatite, and a little scapolite. Again, $4\frac{1}{2}$ miles south-southwest of this locality is a hedenbergite-bearing gabbro, according to Buddington.[34] Admittedly, a great deal of detailed field work must be done before valid generalizations can be attempted, but the occurrence of mutually overlapping mineral associations in an undeformed dike, in a sheared zone, and in "strongly gneissic" gabbroic rock would suggest some overlap of flowing and breaking in the wall rock.

[32] Specimen 26, pp. 36–37 of ftn. 1 (1939).

[33] P. 184 of ftn. 24 (1927).

[34] *Ibid.*, p. 37, specimen 30.

CHILLED BORDER VERSUS DEFORMED BORDER OF ANORTHOSITE

The writer has expressed doubt as to whether there is compelling evidence that the border of the Adirondack anorthosite shows effects of chilling.[35] As Buddington reaffirms his belief in the chilled nature of the anorthosite border,[36] some comments may be submitted on the nature of the evidence. There may be disagreement on what constitutes a chilled border. The writer was of the opinion that a chilled-border rock should be a fine-grained rock, as measured by absolute standards—say average grain size under 1 millimeter. It appears, however, that the rock is considered a chilled phase by Buddington because it is thought that the original rock that crystallized in this border zone was finer grained than the core rocks of the anorthosite massif. The actually exposed border rocks are medium grained and, in almost all outcrops, strongly foliated. According to Buddington,[37] the rocks of the border facies were originally coarser grained. He assumes an original porphyritic rock with plagioclase phenocrysts, 2–5 cm. in length, and a groundmass, "probably not much coarser than 2–6 mm. and certainly not coarser than a centimeter."[38] This may be so; but the reader will ask how Buddington knows that the now crushed and deformed portions of the rock—the bulk of the so-called "Whiteface facies"—with their present medium grain (5–8 mm. in six-

[35] Pp. 391–92 of ftn. 2 (1931).

[36] Pp. 46–48, 325, and 326 of ftn. 1 (1939).

[37] *Ibid.*, p. 46.

[38] *Ibid.*, p. 46.

teen specimens in the writer's collection) have not been derived from anorthosite that contained plagioclases appreciably larger than the few that happened to survive the crushing? How many plagioclase augen in the deformed border of the Westport dome were measured to derive the basis for the supposed maximum size of the original crystals? Are the uncrushed remnants large or small masses? Within the Marcy anorthosite (the core rock of the massif) there are also portions either devoid of phenocrysts or with few and relatively small ones. Before such a statement can be considered a valid generalization, a comprehensive investigation seems desirable. Perhaps Dr. Buddington has made such a study. If so, the results should be published.

Even if it is assumed that the maximum sizes as stated by Buddington are correct, it may well be questioned whether such coarse grain justifies the designation "chilled." According to Buddington's own estimate, the main anorthosite massif crystallized at a depth of between 3 and 6 miles.[39] If at that depth the border of the large massif froze in response to the chilling temperature of the surrounding Grenville sedimentary rocks, the much smaller outlying lenses of anorthosite, as well as syenite sills, should be expected to have correspondingly finer-grained border zones. This should be a common feature. Although the writer does not possess detailed quantitative data concerning grain sizes in Adirondack syenites, his general experience is that there is no such widespread development of fine-grained borders—at least not in the eastern Adirondacks. Furthermore, it should be borne in mind that the grain size of intrusive rocks is the result of several factors of crystallization other than the hypothetical temperature of the wall rock. The literature contains references to relatively large masses of fine-grained rocks and also to small massifs of medium-grained rocks, in which no marginal decrease of grain size is noted, even though many such rocks have probably crystallized at much shallower depth than the Adirondack anorthosite.

The writer has great difficulty following Buddington's discussion of the deformation of the anorthosite border. If he understands the description correctly, Buddington assumes[40] that the primary, porphyritic rock, after complete consolidation, has been mechanically crushed and that, after recrystallization, this rock became the present Whiteface facies of the anorthosite. No photographic or other illustrations are given to show the nature of the "inherited somewhat deformed ophitic pattern" in the rock, nor are localities cited. At any rate, a primary rock fabric must have formed in order to be subsequently blurred and almost destroyed by mechanical deformation of the border. But later on, it is stated[41] that in this area (Buddington's Zone 4 and 4-*a*) there was "little or no crushing of the rocks" and that in this zone the "stress" was at a minimum. The term "stress" is unfortunately not defined; although it may conceivably have reference to both compressive stress and deformation, like the somewhat ambiguous use of the petrographic term "stress mineral," it seems more probable that it has been used to denote deformation or strain. According to Buddington's own estimate of depth of intrusion, the confining compressive crustal stress was at a maximum in the eastern anorthosite region. Jeffries and Archer's statement is acknowledged[42] that "excessive deforma-

[39] *Ibid.*, pp. 328, 333.

[40] *Ibid.*, pp. 46–47.

[41] *Ibid.*, pp. 252, 329.

[42] *Ibid.*, p. 330.

tion may result in a finer grain even though the temperature was high," but this experience in metallurgy has apparently not been considered of significance in interpreting the chilled border. The reasons for this are not clear, the more so as Buddington himself writes:

> Most of the rock in the Adirondacks shows both the phenomena of crushing and that of concomitant recrystallization and, insofar as the evidence inherent in many specimens is concerned, might be called either a protoclastic structure formed during the flowage of an incompletely consolidated magma or a crystalloblastic structure formed by subsequent crushing and associated recrystallization in a solid rock.[43]

The difficulty in following Buddington's conception is further enhanced by his failure to define, or describe in examples, what he considers evidence of "secondary" deformation. In the report on Willsboro quadrangle one reads[44] that the secondary foliation "is in general conformable with the primary foliation." No circumstantial evidence is cited for the existence of two plane structures, observed in one and the same ledge, specimen, or thin section of rocks, striking and dipping in identical directions, yet differing in age by "many millions of years." Does "secondary foliation" mean that in the foliation planes minerals of different periods of crystallization have been found; for example, that hornblende, biotite, or both, have completed their crystallization after that of pyroxene? Or that garnets were observed to penetrate other mafic minerals? If so, how do we know that these minerals have not metasomatically replaced the earlier minerals? Are there microscopically or megascopically visible strike faults that displace, say, crosscutting dikes or veinlets? On another page in the same report[45] one reads that the foliation of anorthosite, where it encloses concordant Grenville skarn layers, is "an initial primary magmatic structure." It would be most desirable to know the difference, as found under the microscope, between rocks of these different histories, but the writer fails to find this information. In the treatment of "dynamothermal metamorphism," it is not made clear whether this process is considered to cause mechanical dislocations in the previously crystallized rocks or chemical and mineralogical changes (i.e., recrystallization or neomineralization with, or without, metasomatic replacement), or whether evidence of both processes has been observed in the region.

One can hardly leave this subject without pausing, also, to consider another implication. If we assume that the Whiteface facies crystallized near the then chilly wall rocks, why is this rock now found miles away from what is considered the general contact? Distances of several miles are obtained on the Lake Placid, Ausable, or Schroon Lake quadrangles. Although the lack of structural data on these sheets makes it impossible to draw cross sections across what might be the prevailing dips of the anorthositic rocks, even a generous allowance for gentle dips yields stratigraphic thicknesses of anorthosite exceeding a mile and probably amounting to far more. Several of these Whiteface anorthosite areas are separated from the main Grenville contact by zones of supposedly massive Marcy anorthosite. If the supposition is made, then, that the roof of the anorthosite massif was "dimpled"[46] here and that the Whiteface anorthosite marks "lows" in the undulating surface,

[43] *Ibid.*, p. 252.

[44] P. 57 of ftn. 6 (1941).

[45] *Ibid.*, p. 37.

[46] Buddington, p. 204 of ftn. 1 (1939).

it is necessary to support such statements with reasonably detailed strike and dip maps of critical areas. Without such, the statement lacks evidence.

That the Whiteface facies is more mafic than the core of the anorthosite massif is readily admitted. In view of the fact that neither Buddington nor N. L. Bowen[47] claim that a chilled border must be more mafic than the core, the writer attaches little significance to this point. As Buddington points out repeatedly,[48] anorthosite with digested Grenville skarn rocks is liable to become more mafic in composition. The higher proportion of these minerals in the border zone, therefore, could also be interpreted as due to assimilation of sedimentary material and hardly seems effective evidence of the chilled nature of the border.

IS THE ANORTHOSITE MASSIF FLOORED?

Miller[49] criticizes the writer's suggestion that the massif may be a north-northeastward dipping lens rather than a "steep-dipping domical sill." In the absence of a diagram by Miller, the writer is unable to visualize a domical sill. Is it a sill on the surface of which a dome protrudes? If the sill is dipping steeply, does the axis of the dome lie about horizontally, or does the dome also exhibit a steep axis? Or is a domical sill a body which has a floor but a domical upper surface? If so, it might be called a "laccolith," provided its dome surface strikes and dips parallel to the roof rocks and provided there is some evidence of a floor.

It is readily admitted that there is no compelling evidence of a floor. It is inconceivable that natural exposures in a region of only 5,000 feet relief should exhibit the actual floor. In the absence of an exposed floor, it seems in order to examine indirect lines of evidence from which the possible shape of the massif might be inferred—at least in local areas. One of these features which, in the writer's opinion, merits consideration, is the remarkable fact that on the southwestern border the anorthosite is accompanied by a belt of northeastward- or northward-dipping Grenville sediments. True, the northerly dips do not continue all the way into, or under, the anorthosite. If they did, this fact would no doubt have been discovered long ago and the appropriate conclusions drawn. Instead, in the Newcomb quadrangle, from which more data are available to the writer than from adjoining sheets, a narrow zone of steep dips follows as one goes northward; and the next few miles, before the anorthosite contact is reached, are underlain by thick, platelike masses of syenite, which dip at high angles to the south or southwest, although local deviations, called "swirls" by Miller, are also noted. The amount of incorporated Grenville material may be larger than originally considered by the writer, especially where micaceous gneisses have been intruded by syenite.[50] The zone, which was traced for about 15 miles, shows a general wedge-shaped cross section;[51] and it was argued that such a structure might be expected to develop in front of an intrusion, rising from the northeast obliquely upward to the south-

[47] "The Later Stages of the Evolution of the Igneous Rocks," *Jour. Geol.*, Vol. XXIII, Suppl. (1915), p. 14.

[48] Pp. 43, 45, 47 of ftn. 1 (1939); pp. 37–39 of ftn. 6 (1941).

[49] Pp. 261 and 264 of ftn. 4 (1943).

[50] The writer is under obligation to Mr. Robert Black, of Syracuse University, for having kindly shown him several significant outcrops of such rocks.

[51] Fig. 6,*b*, p. 328, and Pl. 13 of ftn. 2 (1931).

west. Buddington[52] objects to this inference because

> a fan structure of foliation, similar to that described from the Newcomb quadrangle and on an equally large scale, is also developed in rocks which are predominantly metamorphosed (Grenville) in the Grenville belt of the northwest Adirondacks and cannot therefore be taken as proof of formation in a flowing magma.

The writer fails to see the logic of this argument; but perhaps Buddington has misunderstood the writer, and it may be best to restate the interpretation as follows: The wedgelike structure of the "fan zone" is considered significant because it resembles in its general features and even in details (as, for instance, in the orientation of fold axes and lineation) the structure of highly deformed rocks found in front of overturned folds, upthrust fault blocks, and thrust sheets. If the anorthosite border were the upper portion of an overturned fold, there would be no question but that the layers would bend under, even though actual surface exposures may conceal the trough of such an overturned syncline. As the arrangement of planar and linear structures in this zone suggests strongly that the anorthosite has, indeed, exerted a strong compressive force from the northeast to the southwest, the writer's interpretation seems to have a fair degree of mechanical probability on its side.[53]

Miller points out that the strike of the "axis" or wedge-shaped zone is not exactly parallel to the anorthosite border. Is this lack of parallelism significant? If a thrust block were substituted for the supposed oblique lens of anorthosite, nothing unusual would be seen in a divergence in strike between the overthrust front and the overridden strata. A series of recent structure maps published by the Canadian Geological Survey for the province of Alberta, for instance, affords many good examples of this kind. Furthermore, the Grenville strata commonly exhibit lenticular belts rather than throughgoing zones of constant thickness. The reasons for this difference are well known. Dr. Miller also objects to the position of the axis as drawn in the west-central portion of Schroon Lake quadrangle. Here the tentative trend of the zone was drawn with dashes, as the writer's own field work in this quadrangle was not so detailed that he could be sure of the exact position. It may well be that the axis of the zone extends from west of Hewitt Pond, on the western quadrangle border, in an east-southeasterly direction to the vicinity of Beech Hill. However, as the writer's strike and dip measurements did not everywhere agree with Miller's, the axis was tentatively drawn $3\frac{1}{2}$ miles farther south, because here the continuous Grenville

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[52] Pp. 202–3 of ftn. 1 (1939).

[53] For comparison of cross sections through identical types of structure, either in front of intrusions or in front of thrust blocks or overturned folds, see the following: J. B. Hadley, "Stratigraphy, Structure, and Petrology of the Mount Cube Area, New Hampshire," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), Pl. 4, secs. *D–D'*, *E–E'*; D. Trumpy, "Pre-Cretaceous of Colombia," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), Fig. 6, p. 1298; C. C. Albritton, Jr., "Stratigraphy and Structure of the Malone Mountains, Texas," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), Pl. 1, sec. *G–H;* R. H. Beckwith, "Structure of the Southwest Margin of the Laramie Basin," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), Pl. 1, secs. *F–F'*, *E–E'*, *B–B'*; W. H. Bucher, *The Deformation of the Earth's Crust*, (Princeton: Princeton University Press, 1933), p. 167, Fig. 40, p. 167, and Fig. 55, p. 194, profile *d* (Rosswald); Albert Heim, *Geologie der Schweiz*, Vol. II, 1st half (Leipzig: C. H. Tauchnitz, 1921), Pl. 17 B, opp. p. 364, second and third profiles from bottom; Pl. 18, opp. p. 384, top profile (Mürtschenstock), second profile (Deyenstock), and bottom profile (Fluhbrig); Pl. 21, opp. p. 476, top profile (Vallée de la Gruyère to Au Gros); and many others.

marble belt passes through, so that this position seemed to agree better with the information available farther west. It may also be that there are two subparallel structural synclines and that the one in Schroon Lake quadrangle takes over where a more southerly in Newcomb quadrangle leaves off. Nevertheless, in the west-central portion of Schroon Lake quadrangle, southerly and southwesterly dips seem to prevail north of a line extending from the west-southwest base, or spur, of Bailey Hill in an east-southeasterly direction to the village of Schroon Lake. A structural syncline also passes through the east shore of Cheney Pond, southeast of hill 1970′. Several gabbro bodies complicate the structure pattern.

With regard to strike and dip symbols on several quadrangle maps mentioned by Miller,[54] reference to the Long Lake and Lake Placid quadrangles shows that quite a number of symbols are printed, but unfortunately without dip figures. Even though the text accompanying such maps may state that "primary foliation dips vary in amount and direction, but that relatively steep dips greatly predominate," other persons using the maps cannot tell whether a given symbol stands for an angle of 1°, 89°, or any intermediate value. So long as figures are not added, no cross sections can be drawn through such maps, which is particularly to be regretted, for, as long as no reliable structure cross sections can be drawn, the configuration of the upper surface of the anorthosite massif and the shapes of many other intrusives in the region remain obscure (p. 311). On the Schroon Lake map the writer counts one hundred and thirty-four symbols, fifteen of which have dip figures. That Miller's more recent Lyon Mountain quadrangle does have dip figures was expressly acknowledged.[55]

On the Long Lake quadrangle, by Cushing, there appear both the customary strike and dip symbols, without figures, and straight lines without dip arrows, without explanation in the legend. Their directions commonly deviate from the strike lines of foliation. That Cushing has confounded linear and planar flow structures in these rocks was pointed out by the writer.[56]

Miller criticizes the lack of dip and pitch figures in the writer's Plate 11 of the 1931 paper. Figures were omitted in this outline map, which comprises over a dozen quadrangle areas, to reduce expenses. To make up for the deficiency, a footnote was inserted on page 324 informing the reader that photostat copies of each quadrangle, with the complete dip and strike figures, were procurable at cost from the writer. Several geologists have availed themselves of the offer.

The writer's comments in 1931 on the relative scarcity of figures on maps were an expression of regret, and not destructive criticism of other workers. There is every reason to believe that future quadrangle maps of the Adirondacks will contain more figures, as do the more recently published ones. Structural geology of intrusive rocks is a relatively recent branch of our science, and it will take time before it receives as much attention in the field and in publications as do other features of the rocks.

GENERAL SHAPE OF ANORTHOSITE MASSIF

Concerning the general shape of the large eastern anorthosite massif, there seems to be a misunderstanding between

[54] P. 260 of ftn. 4 (1943).

[55] P. 393 of ftn. 2 (1931).

[56] *Ibid.*, p. 393.

Miller[57] and the writer. The upper portion of a gently dipping lens, as it was represented in the writer's diagram,[58] has, of course, outward-flaring surfaces, at least along the principal contact zones. To that extent, the surface configurations of a dipping lens, of a conical or domical mass, or of a laccolith are identical. If the writer's inferences from the overturned synclinal zone, southwest of the massif, are considered irrelevant by Miller,[59] it is difficult to understand his statement: "The field evidence indicates that the two largest anorthosite bodies are separate intrusions of domical sill or laccolithic shape." What structural field evidence supports this view? The writer is likewise at a loss to understand Buddington's interpretation. After discussing the writer's various considerations in favor of a floor and after concluding that none of them are valid,[60] he writes on the next page: "Present evidence suggests that the anorthosite massif may have a laccolithic or a lens-shaped form or some other sheet-like habit, with a strongly dimpled upper surface." Yet on page 332 Buddington expresses his belief that a large granite mass underlies the anorthosite. Its dikes and offshoots have already reached the present surface at innumerable points. It is believed to have acted as an agent in forcing the anorthosite body up. The evidence on which these divergent interpretations rest would seem to call for re-examination.

AGE RELATIONS BETWEEN SYENITE SERIES AND ANORTHOSITE

This important topic will be considered only briefly, as the writer wishes to make further field observations before discussing it in detail. However, two points call for some comments. The first is the structural relationship between syenite bodies and anorthosite, and particularly the nature of their contacts; the second is the interpretation of andesine augen in syenitic rocks that surround the anorthosite massif.

If the contacts of a dike form an angle with the foliation, lineation, or both, of the surrounding igneous rock, we say the dike is discordant and younger than the wall rock. In general, the meaning of the word "younger" is clear: it is assumed that the wall rock was thoroughly consolidated, and had fractured, before the later dike was intruded. If all the syenite dikes in the anorthosite region were of this nature, that is if they were all discordant bodies, with sharp contacts across the structure of the anorthosite, their relative age would be no problem. In reality, however, these discordant, sharp dikes are rare; in the writer's experience, they are extremely rare. The great majority of contacts which the writer has seen show a zone, many feet wide, of strongly foliated anorthosite and syenite. Contact planes and foliation planes are usually parallel to each other. In such cases the age relations are more obscure. The anorthosite may have been foliated and completely solidified before the arrival of the syenite. The contact would merely be the trace of a fracture, parallel to the foliation of the anorthosite. But it could also be assumed that the wall rock still contained enough of intergranular liquid to be in shearing motion, in layers parallel to the foliation planes, and that the dike magma had enough force of intrusion to wedge in along one of the shear planes and drive the opposite walls aside. Concordant granite and pegmatite dikes in injected gneisses are

[57] P. 258 of ftn. 4 (1943).
[58] Fig. 6, p. 328, of ftn. 2 (1931).
[59] *Ibid.*, p. 261.
[60] Pp. 202–3 of ftn. 1 (1939).

familiar examples of such relations. The possibility that the anorthosite may still have been soft when the syenite magma was intruded becomes a probability where there are dike contacts along which the foliation of the anorthosite is bent from a "regional" direction into one parallel to the dike contact; that is, if flexures in the anorthosite accompany the contacts. If it be assumed that the anorthosite was already thoroughly consolidated before intrusion of a dike and yet developed marginal flexures of its foliation, a confining crustal compressive stress must be assumed of such magnitude that the mechanical resistance of the frozen rock and, especially, that the differences in resistance between individual portions of the anorthosite were completely overcome, to produce quasi-homogeneous movements. It is improbable that such flexure zones, after subsequent recrystallization, would happen to produce the same mineral species and mineral associations as were originally precipitated when the anorthosite crystallized from a magma. Andesine augen in flexure zones would have to be sliced to paper-thin films, with minute displacements among them. Surely, the twin lamellae and extinction positions under the microscope would be disturbed. There are many other ways of ascertaining such flexures of a completely solidified rock. What information the writer has on this subject is definitely against the second mode of deformation. If others have important evidence in favor of such movements of completely solidified anorthosite, it should be published in considerable detail for the benefit of future workers in the Adirondacks.

Many authors have referred to Cushing's locality northeast of Tupper Lake, where the anorthosite is crossed by syenite dikes. The writer published a diagram[61] showing the contacts of twenty-six dikes. He agrees with Miller, Alling, and Buddington that there are numerous discordant dikes with sharp contacts; but it is pertinent to add that there are also syenite dikes that are concordant and still others which exhibit marginal flexures of the anorthosite. It is probably significant that the lineation in many dikes is in almost identical directions with the alignment of andesine augen in the surrounding anorthosite. There are dikes with internal foliation and schistosity, and the strike of these planar structure elements is also in directions almost identical with the alignment of the anorthosite augen. If it is held that the syenite dikes are "definitely younger," some qualifications are necessary. It is true that some of the syenite dikes were emplaced in fractured, solidified anorthosite; but, since quite a number of dikes show other contact relations, as just described, the writer would infer that their wall rocks were still soft enough to yield by bending, and the arrangement of andesine augen and foliation within dikes in similar directions could be explained under the assumption that the movements of the anorthosite had not yet entirely ceased when initial fracturing and dike-intrusion began (see above, p. 306). In the writer's opinion, the dikes associated with flexures and other phenomena of mobility in the anorthosite are more critical for the age assignment of the syenite than the discordant dikes, for they alone indicate the earliest stage at which syenite liquid existed. For our conception of the magmatic differentiation in this area the earliest age of these rocks is the important question, not the latest. There is agreement that syenite was also being

[61] Fig. 25, p. 377 of ftn. 2 (1931).

intruded after the anorthosite had essentially frozen into solid rock.

It should be noted that the age relations, as inferred by the writer, are in close harmony with, and may lend support to, Miller's description of the Keene gneiss.[62] The extremely intimate mixing of syenitic facies with anorthositic rock types, as described by him, is to be expected if the anorthosite still contained interstitial liquid at the time the earliest syenite material moved through it, regardless of whether the syenite is considered an independent intrusion, not closely related to the differentiation of a common-parent magma, or a portion of the mother-liquor ("derivative liquid"). Furthermore, in commenting on Stephenson's paper on the Lake Sanford titaniferous iron-ore deposits, Miller expresses precisely the same view as the writer in judging a gabbro body "practically contemporaneous"[63] with anorthosite because this rock intrudes anorthosite on one side but grades into gabbroic anorthosite elsewhere. The writer prefers to postpone further comments on this point until he has had an opportunity of making detailed field and laboratory studies of these mixed rocks.

Concerning the occurrence of andesine augen in syenitic rocks, in the vicinity of the anorthosite massif, there is agreement that they are found both in dikes and in larger masses of syenite. If these scattered augen are the result of mechanical cracking and comminution of completely solidified anorthosite, should one not expect to find all kinds of anorthositic fragments, from large blocks down to microscopically small crystal fragments? In the writer's experience, neither large blocks, such as Buddington describes from Jenkins Mountain,[64] nor microscopic fragments are nearly so common as the single andesine crystals, of the same size limits as those in anorthosite. This discrepancy needs an explanation, unless the observations, now available, are found to be in error. For this reason the writer wishes to make further studies in the field before discussing other aspects of this problem.

There are many specific, and more general, problems brought out in Alling's, Buddington's, and Miller's discussions; but it is impracticable to extend the present paper without re-examination of certain critical areas in the Adirondacks. Future discussion will undoubtedly be aided if Alling's advice to define terms more precisely is heeded.[65] The writer cannot find a definition of Buddington's "regional dynamo-thermal metamorphism." The phenomena that would fall under it are themselves so numerous, complicated, and open to so many varied interpretations that it is most improbable that such an all-inclusive term will be used in the same sense by various workers. A similar difficulty is experienced with "orogenic forces of mountain-building intensity."[66] Almost anything may be included in it, from litpar-lit injection to volcanism and normal faulting. Not knowing where the emphasis may be, the reader is in danger of substituting his own interpretation, and misunderstanding in the literature may be the result. If ages of structures or rocks are defined as "primary" or "secondary," it would be helpful if the author would explain how he uses these terms, that is, with reference to which other structural feature, or period of

[62] Miller, "Adirondack Anorthosite," *Bull. Geol. Soc. Amer.*, Vol. XXIX (1918), pp. 437–51.

[63] P. 262 of ftn. 4 (1943).

[64] P. 123 of ftn. 1 (1939).

[65] P. 237 of ftn. 3 (1932).

[66] P. 305 of ftn. 1 (1939).

complete solidification of which other rock, the age of another structure or rock is judged. A statement by Buddington reads:

> Most of the rock in the Adirondacks shows both the phenomena of crushing and that of concomitant recrystallization and, insofar as the evidence inherent in many specimens is concerned, might be called either a protoclastic structure formed during the flowage of an incompletely consolidated magma or a crystalloblastic structure formed by subsequent crushing and associated recrystallization in a solid rock. The writer believes, however, that a regional study of the origin of the foliation as it is now found, with respect to the character of crushing and recrystallization and the development of mineralogical facies, indicates that it originated through deformation, flow, recrystallization and reconstitution in the solid state.[67]

This is so broad and inclusive a statement that it is impossible to disagree. The manifestations of a very large magma reservoir, in which various fractions crystallize out, and from which residual fluids are being expelled, at different times, in different directions, and in variable amounts, might easily produce some such variety of features, structures, and age differences of minerals. Yet subsequent passages and discussions leave the present writer in doubt as to how far the agreement goes.

In offering these comments the writer is, of course, keenly aware that terms employed by him are equally open to criticism and that much improvement is needed. He will endeavor to remedy some of the causes in the future.

It is a fair probability that much greater agreement on many problems could be achieved if all Adirondack geologists could arrange a comprehensive joint field excursion. So long as this seems impossible, it would materially facilitate the work if one geologist would be enabled to visit key localities of another, so that all can be sure that they have seen the same rocks. Accurate location of all important exposures, therefore, seems one method of facilitating our joint labors in this important region.

ACKNOWLEDGMENTS.—The writer is indebted to Drs. N. L. Bowen and C. Lochman for reading the manuscript of this paper and for pointing out several inaccuracies and suggesting improvements.

[67] *Ibid.*, p. 252.

ORDOVICIAN BENTONITES IN TENNESSEE AND ADJACENT STATES

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ABSTRACT

Altogether, fifteen beds of bentonite have been recognized in the Ordovician from the lower Stones River group to the lower Maysville. Formerly, it was thought that the Ordovician volcanic activity started in Lowville time and ceased in Trenton time, but beds of bentonite in this area have been recognized stratigraphically far below and above these horizons. Two thick beds of bentonite can be recognized with certainty throughout the western half of the Appalachian Valley from northwestern Georgia and northeastern Alabama to southwestern Virginia.

INTRODUCTION

Although W. A. Nelson[1] described the occurrence and distribution of a single volcanic ash bed in the Ordovician of Tennessee, Kentucky, and Alabama, and R. A. Laurence[2] and Roy Caldwell[3] recognized additional beds in 1936, the total stratigraphic and areal distribution of all the Ordovician bentonites in Tennessee has not previously been fully recognized. Recent dam-site exploration and deep excavations by the Tennessee Valley Authority for the construction of Chickamauga Dam on the Tennessee River, 7 miles upstream from Chattanooga, Tennessee, have revealed a total of fifteen bentonite beds in the Ordovician. Geologic investigations and mapping of the reservoir of Chickamauga, Watts Bar, Norris, and other dam sites have enabled the writers to study the occurrence of the Ordovician bentonites over a large part of Tennessee (Fig. 1). All the Ordovician formations above the Chazy, except the lowest Murfreesboro and upper Sequatchie formations, were fully exposed in large deep excavations and were penetrated by hundreds of large- and small-diameter core-drill holes at Chickamauga Dam. G. M. Kay,[4] R. L. Bates,[5] R. R. Rosenkrans,[6] C. S. Ross,[7] and others have done considerable work on the Ordovician bentonites in Virginia; but there has been little recent work in Tennessee. Insufficient studies have been made near the Tennessee-Virginia line to definitely correlate the beds of bentonite in both states, but the writers believe that detailed work will reveal that several of the beds are the same.

PREVIOUS CLASSIFICATION

Ordovician formations underlie large areas in central Tennessee and crop out in numerous long but narrow belts in the Appalachian Valley in Tennessee (Fig. 1). The Ordovician strata have been given various formation names by many emi-

[1] "Volcanic Ash Bed in the Ordovician of Tennessee, Kentucky, and Alabama," *Bull. Geol. Soc. Amer.*, Vol. XXXIII (1922), pp. 605–16.

[2] "Ordovician Bentonite of the Tennessee Valley Area," *Tenn. Valley Authority, Div. Geol., Bull. 5* (1936), pp. 42–44.

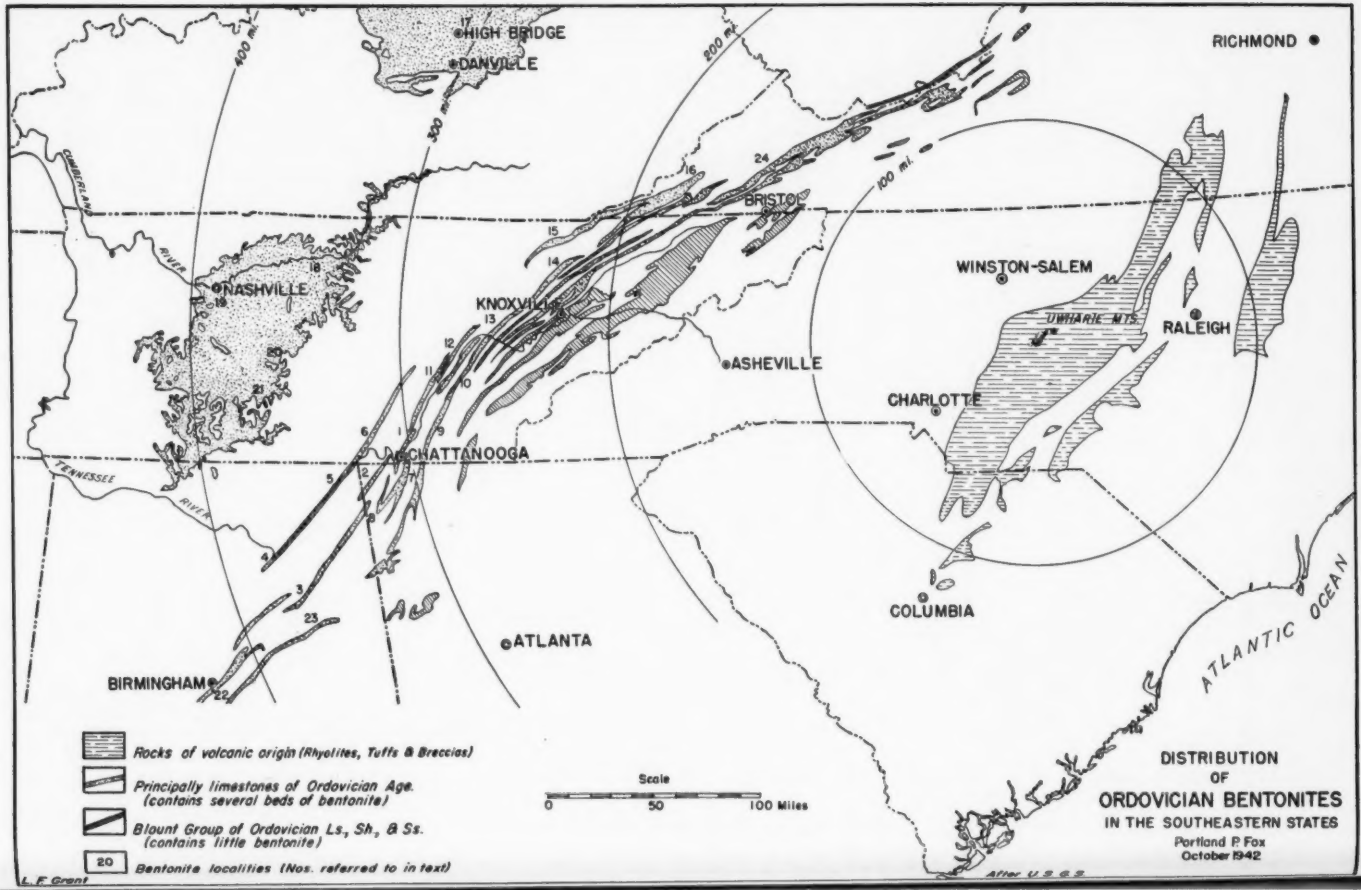
[3] "Metabentonites in Chattanooga Region," *Tenn. Valley Authority, Div. Geol., Bull. 5* (1936), pp. 44–47.

[4] "Distribution of Ordovician Altered Volcanic Materials and Related Clays," *Bull. Geol. Soc. Amer.*, Vol. XLVI (1935), pp. 225–44.

[5] "Geology of Powell Valley in Northeastern Lee County, Virginia," *Va. Geol. Surv. Bull. 51-B* (1939), pp. 49–54.

[6] "Stratigraphy of Ordovician Bentonite Beds in Southwest Virginia," *Va. Geol. Surv. Bull. 46-I* (1936), pp. 87–111.

[7] "Altered Paleozoic Volcanic Materials and Their Recognition," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XII (1928), p. 157.


HIGH BRIDGE
DANVILLE
RICHMOND
400 mi.
300 mi.
200 mi.
100 mi.
CUMBERLAND RIVER
NASHVILLE
KNOXVILLE
BRISTOL
WINSTON-SALEM
UWHARRIE MTS.
RALEIGH
ASHEVILLE
CHARLOTTE
CHATTANOOGA
TENNESSEE RIVER
COLUMBIA
ATLANTIC OCEAN
ATLANTA
BIRMINGHAM
Rocks of volcanic origin (Rhyolites, Tuffs & Breccias)
Principally limestones of Ordovician Age. (contains several beds of bentonite)
Blount Group of Ordovician Ls., Sh., & Ss. (contains little bentonite)
20 Bentonite localities (Nos. referred to in text)
Scale
0 50 100 Miles
DISTRIBUTION OF ORDOVICIAN BENTONITES IN THE SOUTHEASTERN STATES
Portland P. Fox
October 1942
L. F. Grant
After U.S.G.S.

nent geologists. The Alabama, Georgia, and Tennessee portions of the Appalachian Valley are largely covered by several folios of the United States Geological Survey; but in many of these folios the entire Ordovician was frequently mapped as the Chickamauga limestone, with few details on lithology and fossils. R. S. Bassler[8] has published the only comprehensive report on the Ordovician in Tennessee, and the formation names used in this article are modified slightly after Bassler's classification. There is a great variation from west to east in the Ordovician strata in the Appalachian Valley in Tennessee; however, the Ordovician formations occurring along the western portion of the Valley can be correlated with the Ordovician formations of the Central Basin of Tennessee. The Ordovician formations in the eastern side of the Valley in Tennessee are correlated, with considerable uncertainty, with the formations along the western side. Most of the bentonite occurs in the western half of the Appalachian Valley. The classification proposed by Byron N. Cooper and Chilton E. Prouty[9] for the lower Middle Ordovician in Virginia could perhaps be well applied in Tennessee, but that is beyond the scope of this article.

DESCRIPTION OF THE FORMATIONS AND BENTONITE BEDS

GENERAL STATEMENT

Briefly the Ordovician formations will be described in their proper places along with the beds of bentonite. For convenience, the beds of bentonite have been designated B-1, B-2, etc., from the oldest to the youngest (Fig. 2). The oldest bed, B-1, occurs in the lower Murfreesboro formation, which is the lower Chazy.

MURFREESBORO FORMATION

Disconformably overlying the Beekmantown group of limestones, which they closely resemble, is a zone of from 200 to 500 feet of vaughanite, thin-bedded limestone, and silty, mottled red and green dolomite. The lowest member of the formation consists of $7\frac{1}{2}$ feet of conglomerate, composed of large boulders of Beekmantown dolomite and angular chert cemented with silty, light-gray dolomite. Bed B-1, which is described below, overlies the conglomerate. Bed B-1 is followed by from 50 to 135 feet of interbedded vaughanite and mottled red and greenish-gray dolomite. This, in turn, is followed by from 50 to 175 feet of vaughanite. From 100 to 200 feet of thin-bedded, dense, gray limestone composes the upper part of the formation.

Fossils are not numerous in the Murfreesboro; but weathered pieces of the vaughanite frequently contain *Tetradium syringoporoides* Ulrich, and some of the muddy slabs show poor specimens of *Dinorthis* sp. (*D. deflecta?*).

BED B-1

Good exposures of either the Murfreesboro formation or the bentonite are not common in eastern Tennessee, since the formation generally occupies the valleys. However, along the right bluff of the Tennessee River at the Riverview Country Club in North Chattanooga, and at the east side of the village of Washington, Rhea County, Tennessee, 3 feet of weathered bentonite and shale

[8] "Stratigraphy of the Central Basin of Tennessee," *Tenn. Div. Geol. Bull. 38* (1934), pp. 605–16.

[9] "Stratigraphy of the Lower Middle Ordovician of Tazewell County, Virginia," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 819–86.

COLUMNAR SECTIONS
OF
ORDOVICIAN BENTONITES
PORTLAND P. FOX
GENERAL SECTION FOR CHICKAMAUGA DAM AND VICINITY. (Except bed No. 1)
LARGE SCALE SECTION FOR CHICKAMAUGA DAM
Section 3 miles northeast of Dayton, Rhea Co., Tennessee
Section 5 miles northeast of Decatur, Meigs Co., Tennessee
Scaled section for the Norris Reservoir Area
Section 4 miles north of Dandridge, Jefferson Co., Tennessee
Section 4 miles north of Dandridge, Jefferson Co., Tennessee
Section 15 miles southeast of Sharpsville, Bedford Co., Tennessee (Reader)
Section at Hawkins, Tennessee, Smith Co., Tennessee
Section at Turkey Cove, Lee Co., Virginia (Bliss)
EXACT THICKNESS DIFFICULT TO DETERMINE
NORMAL BENTONITE
COARSE SANDY BENTONITE
Scale
Per section No. 1
Per sections No. 2 to No. 11
ORDOVICIAN
LOWER
MIDDLE
UPPER
CHAZYAN
MOHAWKIAN
CINCINNATIAN
STONES RIVER
BLACK RIVER
TRENTON
MAYSVILLE
MARTINSBURG
EDEN
SHERMAN FALLS
LEBANON
LENOIR OR RIDLEY
TYRONE
CARTERS
LOWVILLE
LOWER (MAINLY LIMESTONE)
UPPER (MAINLY CARBON)
RED & GREEN SHALY B.
SEQUATCHIE FORMATION
HERMITAGE
REDMONTOWN
NEWALA
Drawn by L.F. Grant

may be observed. The true character of this bed was not known until several diamond-core drill holes were put down for watertightness investigation in a low saddle of the Watts Bar reservoir in Rhea County, Tennessee. Some unweathered bentonite cores were obtained from these holes; and the material was submitted, for examination, to Ross,[10] of the United States Geological Survey, who is an authority on bentonites. He made the following comments on this material:

> The assemblage of minerals biotite, feldspar, and quartz, the feldspar entirely orthoclase, not plagioclase, is typical of an igneous source. As usual, there is no ash structure visible, but it is evidently a bentonite.

This bed appears to average about 3 feet in thickness, and it appears to be composed largely of hard, gritty, greenish-gray material.

Owing to the rapid thinning of the Murfreesboro formation toward the eastern side of the Appalachian Valley, the occurrence of this bed far to the east side of the Valley is doubtful. Because of its infrequent and poor exposures and its uncertain extent over large areas, this bed will not be of much use in general correlation. However, its presence near the base of the Chazy greatly increases the time span of the Ordovician volcanic activity.

Cooper and Prouty[11] describe a zone of "ash-gray shale" as the second member of their lower Cliffield formation in southwestern Virginia, which may possibly be the equivalent of B-1.

In the Central Basin of Tennessee, a zone of thin-bedded limestone overlying the Murfreesboro formation has been recognized by R. S. Bassler[12] as the Pierce limestone, but it has not been definitely recognized by the writers in the Appalachian Valley. At Chickamauga Dam and at numerous other localities along the western side of the Appalachian Valley in Tennessee, 45 feet of unfossiliferous, red and gray mudstones occur at or near the top of the Murfreesboro (Fig. 2). This zone does not contain any bentonite.

RIDLEY (OR LENOIR?) FORMATION

Overlying the zone of mudstones and underlying the Lebanon limestone at Chickamauga Dam, 340 feet of thin- to thick-bedded, dark bluish-gray limestone was completely exposed (Fig. 2). At this locality the formation contains no bentonite, but a single 3-inch layer of bentonite has been observed in the upper third of the Lenoir formation in a small quarry 10 miles northeast of Sweetwater, Monroe County, Tennessee, A similar bed of bentonite has been noted, by the junior writer, in the Lenoir formation 5 miles east of Knoxville. A number has not been assigned to this bed of bentonite since the correlation of the Lenoir formation with the Ridley formation is in question.

LEBANON FORMATION

Ninety feet of thin-bedded and shaly limestone overlying the Ridley formation and underlying the Carters limestone was completely exposed at Chickamauga Dam. Thin slabs of the limestone are often nearly completely composed of *Hesperorthis tricenaria* Bassler and *Sowerbyella lebanonensis* Bassler. Owing to the formation's shaly character and position between thick-bedded limestone and its abundant fossils, it has been recognized along the western half

[10] Personal communication.

[11] P. 825 of ftn. 9 (1943).

[12] *Ibid.*, pp. 605–16.

of the Appalachian Valley at many places; but nowhere has any bentonite been found within it. The formation as here used appears to correspond approximately to the lower half of the Witten limestone as used by Cooper and Prouty[13] in Virginia.

LOWVILLE FORMATION

This formation at Chickamauga Dam, as elsewhere in this part of the state, consists of two distinct lithologic members. Bassler[14] gave the name "Carters" to the lower and "Tyrone" to the upper member. Along the western half of the Appalachian Valley the Carters consists of 80 feet of medium- to thick-bedded bluish-gray, dense, fossiliferous limestone similar to that at Chickamauga Dam. The thick beds of limestone separating B-2 and B-3 commonly carry numerous fossils such as *Columnaria halli, Buthotrephis inosculata, Camarocladia gracilis,* and *Stromatocerium rugosum.*

Overlying the Carters at Chickamauga Dam and elsewhere along the western half of the Applachian Valley in Tennessee, Georgia, and Alabama, the Tyrone member consists of 40 feet of thin-bedded, greenish-gray, light-gray limestone and red shaly limestone. The distinctive lithologic characteristics of this member readily distinguish the Tyrone from the overlying thick-bedded Hermitage and underlying Carters. Fossils are not abundant; but some of the upper, thin, dense, gray limestones do contain *Tetradium cellulosum.* This member contains several beds of bentonite, and it is separated from the underlying Carters member by bed B-3. The Tyrone in this area is in part equivalent to the Moccasin formation in northeastern Tennessee and southwestern Virginia.

FIG. 3.—Upper portion of the Hermitage formation as exposed in excavation for Chickamauga Dam powerhouse showing B-11 and B-12.

[13] *Ibid.*, p. 842.

[14] Pp. 61–64 of ftn. 8 (1934).

BED B-2

Thirty feet below the top of the Carters in the Chattanooga area, a 1- to 4-inch bed of impure greenish-gray bentonite occurs. A 6-inch bed occupying a similar position has been noted in northeast Hamilton County, Tennessee, a mile west of Georgetown on Tennessee highway No. 58.

FIG. 4.—Bentonite bed B-3 as exposed in a quarry in East Chattanooga, Tennessee. Note layer of chert under the Bentonite about level of man's knees. The bed has been thickened slightly by squeezing.

BED B-3

This bed occurs at the top of the Carters and consists of three types of bentonite (Fig. 4). The lower portion is composed of 3–8 inches of greenish-gray, sandy arkosic bentonite; the middle portion, 15–20 inches of soft, fine-grained, light-gray sheared bentonite; and the upper portion, 8–12 inches of dense, hard, light-gray, impure, shaly bentonite. The coarse lower portion of this bed and an underlying 2- to 4-inch layer of dense, black, fossiliferous chert readily distinguish this bed from all the others. The bed of chert was encountered at 100–200 feet below the top of rock and far below the zone of weathering in numerous core-drill holes at Chickamauga Dam. The coarse bentonite frequently carried a small amount of artesian water, which probably has picked up silica from the bentonite and replaced the calcium carbonate to form the underlying layer of chert. Neither of the other beds of bentonite is underlain by chert, probably because the underlying limestones are impure and less susceptible to replacement and also because the other thick bed of bentonite, B-6, has fine-grained, impervious bentonite at the bottom, its coarse material being near the middle.

Ross examined some of the coarser

material from bed B-3 and described it as follows:

> The material from near Chattanooga is clearly bentonite. I have examined specimen No. 1 most carefully and find that it is unusually arkosic and contains both orthoclase and plagioclase. The plagioclase is very fresh, but there has been a marked alteration of the orthoclase. Some very perfect volcanic ash structures are preserved in the clay portion of this specimen. It is all of the fibrous pumice-like type rather than the rounded bubble type.
>
> Ash structures are commonly not well preserved in the Ordovician bentonites, but those in this material are among the best that I have observed, although good examples are not numerous even in this material.
>
> The grain size is unusually coarse and may indicate a closer proximity to a center of eruption than much of the material we have seen.[15]

BED B-4

Seven feet above bed B-3 is a 1-inch layer of fine-grained, greenish-gray, impure bentonite which has not been definitely recognized outside of the Chattanooga area. At Crudup, Alabama, a thin bed occupies about the same position above bed B-3 (Fig. 2).

BED B-5

This bed occurs $3\frac{1}{2}$ feet above B-4 in the Chattanooga area; and it is similar to B-4 in thickness, grain size, and color. Also, at Crudup, a $7\frac{1}{2}$-inch bed occupies about the same position that it occupies in the Chattanooga area.

BED B-6

Eleven feet above B-5, $2\frac{1}{2}$–4 feet of bentonite occurs slightly above the middle of the Tyrone. This bed consists of three types of bentonite similar to B-3, except that 6 inches of light-gray, fine-grained, impervious bentonite occurs at the base with 6-12 inches of extremely coarse arkosic, micaceous bentonite in the middle of the bed. This is overlain by $1–2\frac{1}{2}$ feet of cream-colored, fine-grained bentonite. This bed is overlain by $\frac{1}{2}$–1 foot of calcareous sandstone. The sandstone above this bed and the coarse micaceous material in the middle make it rather easy to distinguish this bed from B-3. Bed B-8 is quite arkosic, and it could easily be confused with B-6 if the sandstone were absent. In the Chattanooga area, B-6 is underlain by 5 feet of red shaly limestone, which assists greatly in recognizing it.

In 1939 we sent Ross some unweathered cores of bentonite from Chickamauga Dam foundation, and he made the following comments on them:

> I have been interested in examining the three samples of volcanic material from the Chickamauga Dam. No. 1 [arkosic material B-6] has biotite as the most characteristic mineral, but also contains abundant orthoclase, quartz, and a little plagioclase. The quartz is unusually abundant, but seems to be of volcanic rather than detrital origin.
>
> No. 2 [coarser material, B-3] is unusually rich in orthoclase, but contains a little plagioclase. Almost no biotite is present, and but little quartz. The feldspar has been partly replaced by clay material and by a still larger proportion of calcite. Thus it was originally 60 per cent or more crystalline. The grains are largely euhedral, and reach a millimeter in diameter.
>
> No. 3 is largely clay material. [Sample No. 3 was from the fine-grained material near the top of B-3.]
>
> No. 2 is most conspicuously coarse-grained, but No. 1 is also rather coarse-grained, and although No. 3 retains the original structure poorly, it, too, was probably made up of glass grains up to a millimeter in diameter.
>
> This coarseness of grain is the most interesting thing, as it suggests closer proximity to a source than heretofore observed.[16]

BED B-7

In the Chattanooga area this bed occurs $5\frac{1}{2}$ feet above B-6, and it is simi-

[15] Personal communication, 1935.

[16] Personal communication, 1939.

lar to B-4 and B-5 in thickness, color, and grain size. It has not been definitely recognized outside this area, but a bed in about the same position occurs in the section near Decatur, Tennessee (Fig. 2).

BED B-8

This bed occurs 5 feet above B-7 and is quite similar to B-3 and B-6. In several core holes at Chickamauga Dam it consisted entirely of 4–8 inches of green, coarse, micaceous, arkosic bentonite; but in other holes the arkosic material was completely absent and 1–2½ feet of soft, cream-colored, fine-grained bentonite represented the bed. This bed sometimes reaches a thickness equal to B-3 and B-6, but its average thickness is only about one-third that of B-3 and B-6. It occurs near the top of the Tyrone and appears to be absent throughout most of eastern Tennessee.

BED B-9

At the top of the Tyrone and the bottom of the Trenton (Hermitage) at Chickamauga Dam, a 1- to 2-inch layer of fine-grained, gray bentonite may be seen. This is absent in the Avondale sections of Chattanooga a few miles to the southwest. The Trenton and Lowville are conformable in the Chattanooga area, although the oldest Trenton is not everywhere present in this general region.

TRENTON GROUP

Throughout eastern Tennessee the Trenton is represented by two formations (Fig. 3). Bassler[17] assigned the names "Curdsville," "Hermitage," "Jessamine," "Bigby," "Cannon," and "Catheys" to the six formations composing the Trenton in the Central Basin. The Hermitage is represented at Chickamauga Dam by 65 feet of thick-bedded, argillaceous, nodular, fossiliferous, dark bluish-gray limestone. *Dalmenella fertilis*, *Lichenaria globularis*, *Rhynchotrema increbescens*, and *Hindia sphaeroidalis* are the common fossil forms in the Hermitage along the western side of the valley in Tennessee.

BED B-10

Thirteen feet above B-9, at the base of the Hermitage, lies 3–5 inches of soft, cream-colored, fine-grained bentonite, which is overlain by an equal thickness of dark-gray, calcareous shale.

BED B-11

This bed is slightly thinner than B-10 but is lithologically similar. It lies 28 feet above B-10 and is shown in Figure 3.

BED B-12

Three feet above B-11 there is an extremely thin bed of bentonite. It consists of only ½ inch of greenish-gray, soft, fine-grained material. This bed could not be recognized except in fresh exposures of the bedrock. B-12 lies 13 feet below the top of the Hermitage (Fig. 3).

CANNON FORMATION

At Chickamauga Dam this formation consists of 105 feet of dark-gray, thin- to medium-bedded, dense fetid limestone (Fig. 5). The formation is quite fossiliferous; and *Platystrophia glabosa*, *Hormotoma columbina*, *Lophospira bowdeni*, and *Bellerophon troosti* are common forms. The lower 6 feet of the Cannon is subvaughanitic, and it may represent the Bigby if that formation is present in this area.

BED B-13

Six feet above the base of the Cannon is a 1½-foot layer of shale and bentonite.

[17] P. 71 of ftn. 8 (1934).

The basal inch or two is whitish, fine-grained bentonite with another thinner layer 6 inches higher in the calcareous shale. This layer of shale and bentonite being easily recognized in the drill cores, it was designated as the "key bed" at Chickamauga Dam. Without this bed it would have been almost impossible to have worked out the complicated geologic structure at this dam from drill cores in advance of the excavations.[18]

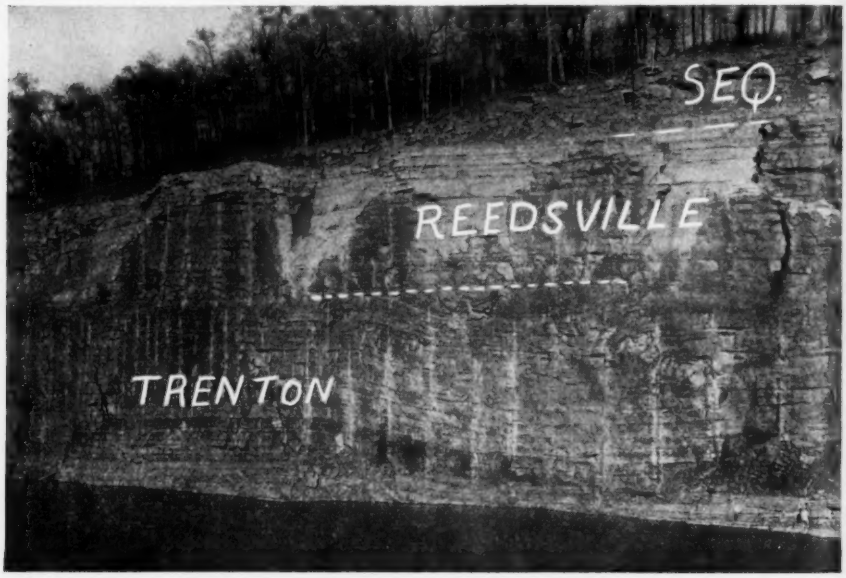


FIG. 5.—Sequatchie, Reedsville, and Upper Trenton limestones as exposed in the main quarry near Chickamauga Dam. Bed B-13 was exposed just below water-level, and B-14 in the upper right-hand corner.

REEDSVILLE FORMATION

Unconformably overlying the Cannon in the Chattanooga area is 40 feet of thin-bedded, dense, greenish-gray, argillaceous limestone of the Reedsville formation (Fig. 5). The formation is not abundantly fossiliferous; but *Plectorthis plicatella*, *Platystrophia ponderosa*, and *Rafinesquina ponderosa* have been collected from its upper portion. No bentonite occurs in the Reedsville in this area. Northeast of Chattanooga this formation becomes progressively thicker, until it reaches several hundred feet in northeastern Tennessee.

SEQUATCHIE FORMATION

Only 40 feet of this formation was exposed in the quarry at Chickamauga Dam (Fig. 5). The lower 15 feet consisted of shaly, red and greenish-gray limestone. The upper portion consisted of 25 feet of thin-bedded, bluish-gray, dense limestone. In the belts east of this one in eastern Tennessee this formation is represented by 200–300 feet of red, muddy, and siliceous limestone immediately under the Brassfield formation.

[18] Portland P. Fox, "Foundation Exploration and Geologic Studies at Chickamauga Dam," *Trans. Amer. Soc. Civil Eng.*, Vol. CVI (1941), pp. 765–79.

BED B-14

Fifteen feet from the base of the Sequatchie, or at the top of the red and greenish-gray, shaly limestone, is a 3- to 4-inch layer of greenish-white, soft, fine-grained bentonite. This bed has not been recognized outside the Chattanooga area; but it is of interest, since it indicates that the volcanic activity did not cease until near the close of the Ordovician.

DESCRIPTION OF BENTONITE LOCALITIES SHOWN ON FIGURE 1

1. *Chickamauga Dam and vicinity.*—Several beds of bentonite of the Lowville, Trenton, and Sequatchie formations are well exposed in numerous quarries from the Avondale section of Chattanooga to the main Chickamauga Dam quarry located 1 mile northeast of the dam (Figs. 4 and 5).

2. *Trenton, Dade County, Georgia.*—The two thick beds of bentonite (B-3 and B-6) in the Lowville are exposed along Highway U.S. 11, 2 miles northeast of Trenton and ½ mile west of that town on a country road. At the latter place, bed B-6 is ususually thick—20 feet—but this is probably due to squeezing.

3. *Crudup, Etowah County, Alabama.*—Six beds of bentonite are exposed in the Trenton and Lowville formations 1½ miles west of Crudup, Etowah County, Alabama (see Fig. 2, Sec. 6).

4. *Guntersville, Marshall County, Alabama.*—Two miles northeast of Guntersville, bed B-3 is seen along Highway U.S. 241. It is slightly less than 2 feet thick.

5. *Bridgeport, Jackson County, Alabama.*—On the right bank of the Tennessee River at the Bridgeport water works, bentonite beds B-3 and B-6 are exposed. Each is about 2½ feet in thickness.

6. *Whitwell, Marion County, Tennessee.*—Five miles east of Whittwell, bentonite bed B-3 is 3 feet thick. Nelson[19] reported a similar thickness at Pikeville, 37 miles northeast of the above location.

7. *Chickamauga National Military Park, Catoosa County, Georgia.*—Bentonite beds B-3 and B-6 are exposed along Highway U.S. 27, 1 mile south of the park and also on Jolly's farm near by. Both appear to average about 3 feet thick.

8. *Harrisburg, Georgia.*—Harry X. Bay and Arthur C. Munyan[20] stated that bed B-3 has a maximum thickness of 8 feet at Harrisburg, Catoosa County, Georgia, while bed B-6 has a thickness of 2 feet at Cooper Heights, Walker County, Georgia.

9. *Georgetown, Bradley County, Tennessee.*—One-half mile west of Georgetown and at several points between this village and the Hiwassee River, beds B-3 and B-6 are well exposed. The thickness is about the same as at Chickamauga Dam.

10. *Decatur, Meigs County, Tennessee.*—Five beds of bentonite are well exposed in a road cut 2 miles northeast of Decatur, Tennessee. The same beds are poorly exposed just east of the courthouse in Decatur (Fig. 2).

11. *Evensville, Rhea County, Tennessee.*—One-half mile west of Evensville bentonite bed B-3 has been worked for a bleaching clay. It has an average thickness of 3 feet.

12. *Watts Bar Dam, Rhea County, Tennessee.*—One mile west of the dam

[19] Pp. 605–15 of ftn. 1 (1922).

[20] "The Bleaching Clays of Georgia," *Ga. Div. Geol. Inf. Circ. No. 6* (1935), pp. 2–3.

on Highway 68, three beds of bentonite are exposed in the upper Lowville. Numerous other exposures occur in the same Ordovician belt a few miles southwest and northeast of the above locality. Thicknesses are similar to the Chickamauga Dam section.

13. *Wheat, Roane County, Tennessee.*—One mile west of Wheat, bed B-3 crops out along Poplar Creek. Also, approximately 5 miles east in another belt, bed B-3 appears in a sharp bend of the Clinch River.

14. *Andersonville, Anderson County, Tennessee.*—At several points northeast and southwest of this village, bed B-3 is exposed.

15. *Caryville, Campbell County, and Speedwell, Claiborne County, Tennessee.*—At several points between Caryville and Speedwell near Highway Tennessee 63, James S. Cullison[21] recognizes that bed B-3 occurs continuously and that bed B-7 occurs intermittently. B-3 and the underlying chert are especially well exposed in the vicinity of Loyston and Sharp's Chapel in Big Valley in Union and Claiborne counties.

16. *Turkey Cove, Lee County, Virginia.*—Bates[22] described two beds of bentonite in the Lowville at this locality. The section described by him is practically identical to the Tennessee sections (Fig. 2). His lowest bed of bentonite is possibly the same bed as our B-3.

17. *High Bridge, Mercer County, Kentucky.*—Bassler[23] describes two beds of bentonite at this locality, the lower being no doubt the same as our bed B-3.

18. *Round Lick Creek Bridge on Highway U.S. 70 N, Smith County, Tennessee.*—Bed B-3 is 12–18 inches thick. Bed B-6 is absent.

19. *Nashville, Tennessee.*—Bassler[24] describes B-3 in his section made from the Vantrease quarry near the Mount Olivert Cemetery as having a thickness of 12 inches, but the senior writer has examined this quarry and measured the bed as only 6 inches thick. In this same quarry two thin (0.1 foot) beds of bentonite occur 30.5 and 53.0 feet below B-3. The one 30.5 feet below B-3 correlated with B-2, but the other bed is unknown elsewhere.

20. *Beech Grove.*—One mile west of this village in Rutherford County, Tennessee, B-3 is exposed along Highway U.S. 41. Its thickness is 1.5± feet.

21. *Singleton, Tennessee.*—See Bassler's[25] section for the locality.

22. *Birmingham, Alabama.*—Charles Butts[26] recognized a bed of bentonite in the Lowville (B-3) and another one at the bottom of the Trenton. Nelson[27] had previously recognized a single bed (B-3) 14–19 inches thick at this locality.

23. *Ragland, Alabama.*—Butts[28] states that "two layers of bentonite, each about one foot thick, occur in the upper 10 to 15 feet of the Little Oak limestone." The Little Oak limestone is probably Trenton in age.

24. *Old Rosedale, Virginia.*—Kay[29] recognized ten beds of bentonite in the Trenton at this locality and also the greatest thickness, 96 inches, for a single bed in the Ordovician.

[21] "Geology of Norris Reservoir" (unpublished manuscript, T.V.A. Geologic Division, 1937).

[22] Pp. 51–52 of ftn. 5 (1939).

[23] Pp. 64–65 of ftn. 8 (1934).

[24] *Ibid.*, p. 22.

[25] *Ibid.*, pp. 64–65.

[26] Butts and Others, "Geology of Alabama." *Geol. Surv. Ala. Spec. Rept. No. 14* (1926), p. 131.

[27] Ftn. 1 (1922).

[28] P. 114 of ftn. 25 (1926).

[29] Ftn. 4 (1935).

PHYSICAL AND CHEMICAL PROPERTIES

A sample of the coarse arkosic bentonite from the middle of B-6 and a typical sample of the fine-grained, whitish bentonite from B-12 were collected from fresh exposures at Chickamauga Dam and analyzed as shown in Tables 1 and 2.

TABLE 1

MECHANICAL ANALYSIS—ROTAP SIEVE METHOD

Sieve No. (U.S. Series)	Sample of Arkosic Bentonite from B-6		Sample of Whitish Bentonite from B-12	
	Grams	Percentage	Grams	Percentage
	Weight Retained			
10	0.5	0.2	1.5	0.4
20	1.5	0.7	1.0	0.7
40	27.5	10.6	3.5	1.6
70	100.5	46.6	13.0	5.0
100	38.5	60.4	15.5	9.0
150	26.0	69.7	22.0	14.7
200	14.5	74.9	46.5	26.7
270	6.0	77.1	16.0	30.8
325	2.0	77.8	13.5	34.3
Weight passing 325 mesh	62.0	22.2	353.4	65.7

The mechanical analyses were made by dispersing the bentonite in distilled water with a propeller-type agitator. The material was then wet-screened and later dry-screened.

Less than 25 per cent of the arkosic bentonite from bed B-6 passed the 325-mesh sieve, but 65.7 per cent of bed B-12 passed the same sieve. By studying the mechanical analyses of a single bed of bentonite over a large area, one might be able to gain additional information on the source of the material.

The higher percentage of Fe_2O_3 in B-6 is probably due to the biotite, which is quite noticeable in this arkosic bentonite. Most of the calcium carbonate in B-12 is due to thin layers of secondary calcite in the bentonite. These analyses are similar to the analyses presented by Nelson[30] of the single bed (B-3) of Ordovician bentonite from Singleton, Tennessee; High Bridge, Kentucky; and Birmingham, Alabama.

SUGGESTIONS AS TO THE SOURCE OF THE BENTONITES

The source of the Ordovician bentonite has always been an interesting topic.

TABLE 2

CHEMICAL ANALYSIS OF BENTONITE*

	Bed B-6	Bed B-12
SiO_2	55.11	47.89
Al_2O_3	18.46	20.63
Fe_2O_3	4.39	1.44
MgO	3.50	3.66
CaO	2.10	7.59
Na_2O	1.27	0.62
K_2O	5.78	5.20
TiO_2	0.25	0.26

* E. C. Houston, analyst.

Nelson[31] first suggested northeastern Kentucky as a possible source, but he later regarded central Virginia as a more probable location of the volcanic activity. Kay[32] suggested western North Carolina as a source for the Ordovician bentonite, because the thickest beds were found in Russell County, Virginia. The writers are quite familiar with the geology of western North Carolina and find little evidence for placing the volcanic center in this region. Arthur Keith[33] has

[30] P. 614 of ftn. 1 (1922).

[31] *Ibid.*, p. 613.

[32] Pp. 243–44 of ftn. 4 (1935).

[33] *U.S. Geol. Surv. Atlas, Roan Mountain Folio 151* (1907).

mapped several small bodies of eruptive rocks in eastern Tennessee and western North Carolina, but they are all interbedded with the Lower Cambrian clastics. The great volcanic activity required for the eruption of the Ordovician bentonites over a long period of time must have had associated with it some extrusives and various pyroclastics. It would seem odd, indeed, that so much of the bentonite and yet none of the extrusives and coarser pyroclastics could be preserved; but farther east we note that a great series of volcanic rocks which has usually been considered pre-Cambrian in age but may be later in age extends across central North Carolina from Virginia into South Carolina (Fig. 1). The writers believe that this group of volcanic rocks offers the best clue for the bentonite source, since it cannot be conclusively proved that their age is greater than the Ordovician. In Union County, North Carolina, this series consists largely of flat-lying, undisturbed acid tuffs. They have been intruded by post-Carboniferous granites and altered considerably near this contact, but their regional metamorphism is so slight as to suggest an age later than pre-Cambrian. The center of this volcanic activity was in the Uwharie Mountains in northwestern Montgomery and southern Randolph counties, central North Carolina. Its location is only 100 miles from the nearest Ordovician outcrops in southwestern Virginia and only 150 miles from Russell County, Virginia, where Kay measured 8 feet on a single bed of bentonite (Fig. 1).

CONCLUSIONS

In eastern Tennessee, bentonite deposition started in early Chazyan and closed in the Cincinnatian epoch. The bentonites in this region have a stratigraphic range of more than 1,500 feet from lower Murfreesboro to lower Sequatchie formations. The lower thick bed of bentonite, B-3, in the Lowville has been recognized in several belts along the western edge of the Appalachian Valley from northeastern Tennessee into Georgia and Alabama. This makes an excellent key bed, and we are of the opinion that future geologists working in the area will learn to appreciate its significance. Many of the thin beds of bentonite apparently have wide extent, but more detailed study and better exposures are necessary before they can be definitely recognized over a wider area.

ACKNOWLEDGMENTS.—The writers have collected these data while engaged as geologists for the Tennessee Valley Authority on the various dams mentioned above. All the geologic investigations have been under the direction of Berlen C. Moneymaker, chief geologist, and Edwin C. Eckel, former chief geologist. The writers are indebted to R. A. Laurence, former geologist of the Tennessee Valley Authority. Many of the fossils have been identified by G. A. Cooper, of the United States National Museum, Charles Butts, and James S. Cullison.

GEOMORPHIC EFFECTS OF INDURATED VENEERS ON GRANITES IN THE SOUTHEASTERN STATES

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ABSTRACT

Indurated surficial veneers are a characteristic phenomenon on exposures of unjointed granite in the southeastern states. They are caused by deposition of iron oxides in the interstices between the partially disaggregated mineral grains of the "sap" rock at the surface of the exposure. Breaching of veneers produces such peculiar erosional forms as weather pits, cavernous boulders, and pedestal rocks.

INTRODUCTION

Indurated surficial veneers or case-hardening effects on granite exposures have long been known in arid regions. A. L. Anderson[1] has described them in detail from the "Cassia City of Rocks" in Idaho, where they have had a manifest effect in determining topographic form; but they have passed virtually unnoticed in humid regions, such as the southeast. However, the Idaho veneers as described by Anderson seem to differ little from a surface-hardening phenomenon which the writer has observed quite commonly on granite exposures throughout the southeastern Piedmont. The southeastern veneers, as well as those in Idaho, appear as dark, hardened layers on the surfaces of unjointed granite exposures that are undergoing granular disintegration; and both seem to result from the deposition of iron oxides in the interstices between the partially disaggregated mineral grains.

ORIGIN OF VENEERS

Speaking of the origin of the Idaho veneers, Anderson says:

> As the outer or near outer surface of the rock masses tend to become desiccated, capillary action draws water from the interior of the rocks and with them dissolved salts and colloids, and as this capillary water in turn evaporates at or near the surface it deposits its borne substances, thus cementing or closing the fractures in the rock and thereby strengthening the disintegrated rock by the process of case-hardening.[2]

However, the presence of durable veneers in a humid climate such as that of the southeastern states fairly well eliminates the possibility of their having been formed by soluble materials. Studies of thin sections made from the southeastern veneers attest the general applicability of Anderson's theory of origin, but the writer believes that it is dominantly the presence of insoluble hematite in the deposited substances which enables the veneer to resist weathering more successfully than the underlying rock.

From its original source in primary iron-bearing minerals, such as hornblende or biotite, the iron would be released as a soluble ferrous compound which would readily be brought to the surface by capillary water. With evaporation of the water the iron would be deposited in an oxidizing environment and would change to the less soluble ferric form, becoming the weather-resistant cementing agent which rebuilds the par-

[1] "Geology and Mineral Resources of Eastern Cassia County, Idaho," *Idaho Bur. of Mines and Geol. Bull. 14* (1931).

[2] *Ibid.*, p. 59.

tially disaggregated granite to make the indurated veneer.

Table 1 shows two chemical analyses, for which the writer is indebted to Dr. W. A. Reid, of the department of chemistry of the North Carolina State College of Agriculture and Engineering in Raleigh. These analyses were made from samples of an indurated veneer and the immediately underlying partially disaggregated granite. Both were obtained

TABLE 1

ANALYSIS OF VENEER AND UNDERLYING ROCK

	Indurated Veneer	Underlying Partially Disaggregated Rock
SiO_2	71.85	73.10
Fe_2O_3	2.41	1.91
Al_2O_3	15.26	16.14
CaO	1.98	1.61
MgO	0.51	0.52
Na_2O	3.55	4.14
K_2O	1.58	1.63
Total	97.14	99.05

from the outcrop shown in Figures 7 and 8.

Two such isolated analyses are admittedly inadequate as a basis for reasoning concerning the genesis of the veneer, but perhaps it will not be amiss to point out certain indicative relations between them. In general, it will be noted that the differences in the percentages of the several components are quite small. However, one would expect this because the principal difference between the veneer and the immediately underlying rock is the presence of the cementing material in the veneer. And the cement comprises only a very small percentage of the veneer, for the primary mineral grains have not been displaced mechanically and the intergranular openings into which the indurating material has been insinuated are exceedingly narrow.

For this reason, minor distinctions in composition may be significant, such as the 0.50 per cent difference between the 2.41 per cent ferric oxide in the veneer and the corresponding 1.91 per cent in the underlying partially dissaggregated rock—a difference of 26 per cent in terms of total iron present in the latter. It will also be noted that calcium, although present in small amount, is likewise proportionately much more abundant in the veneer than in the underlying rock. On the other hand, the more soluble constituents, sodium and potassium, show significantly lower percentages in the veneer than in the underlying rock. Probably they have been alternately deposited during periods of desiccation and leached during wet periods.

Heretofore, veneers have been considered the product of an arid climate, but during a four-year period of field work throughout the southeast the writer has observed them on granite exposures in so many widely scattered localities that he is forced to regard them as a characteristic phenomenon of that humid region also. Perhaps the explanation of their presence in the southeast may be found in the fact that they most frequently occur on exposures which have few or no joints. Such exposures are not readily susceptible to the deep chemical weathering which produces the usual saprolitic surface mantle of the southeast. They resist such deep weathering because they do not have any available avenues for the ingress of surface water by percolation. The only available openings are capillary in size and are probably less frequently occupied by the absorption of surface water than by the rising of ground water to replace that

which evaporates from the sun-heated rock surface.

In this connection it is interesting to note that the unjointed granites of the southeast characteristically weather by granular disintegration, breaking down into a gruss indistinguishable from that which usually results from the weathering of granites in arid climates. Under arid conditions the rock weathers by granular disintegration because of the dearth of rainfall. Under humid conditions the rainfall, although plentiful, has no ready access to the interior of the rock because of the absence of joints.

Once the indurated layer has been established, it is self-perpetuating, growing at the bottom as erosion removes the surface. It is possible that the constant protective cover which it provides for the underlying sap rock may be a factor in determining the regular surfaces which these unjointed granite masses frequently assume. Certainly it is evident in the field that most irregularities in their surfaces occur where some special circumstance has enabled erosion to break through the veneer.

BREACHING OF VENEERS BY RUNNING WATER

In the southeast there are two common agencies which enable erosion to breach the veneer. Both of them expose the veneer in section and thus manifest its otherwise obscure presence. The more easily recognized one is the concentrated flow of surface water in the small drainageways on the surfaces of the exposures. Where such drainageways cut through the veneer, they usually undercut it by widening themselves in the softer underlying sap rock. Figure 1 shows a photograph of such a drainageway. It will be noted that the veneer has been breached and is prominently displayed as an overhanging lip where it has been undercut.

ORIGIN OF WEATHER PITS

Although the other breaching agency is less readily identified, the depressions

FIG. 1.—Minor drainageway in an exposure of unjointed granite showing indurated veneer as an overhanging lip where it has been undercut. Near Trap Hill, Wilkes County, North Carolina.

which result from its activity are of very uniform character and have been described in the literature as "weather pits." They have been discussed in excellent detail by L. L. Smith[3] and more

[3] "Weather Pits in Granite of the Southern Piedmont," *Jour. Geomorph.*, Vol. IV (1941), pp. 117–27.

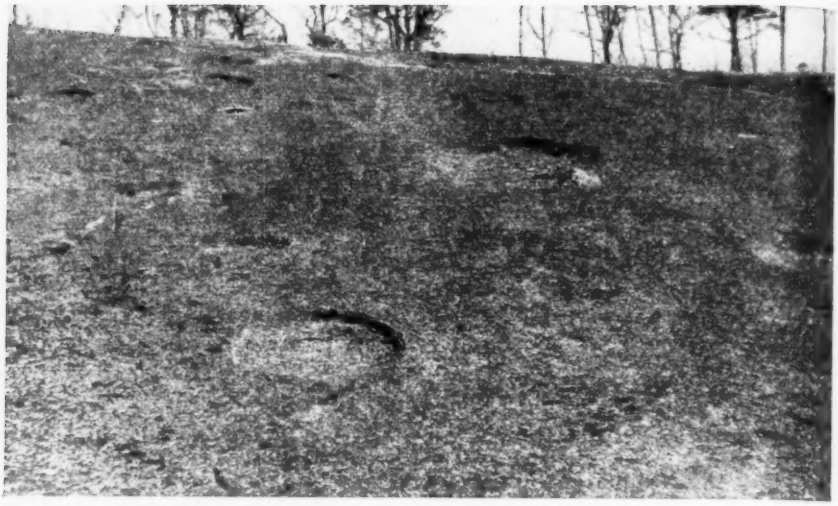

FIG. 2.—Small weather pits in granite, near Stone Mountain in Wilkes County, North Carolina

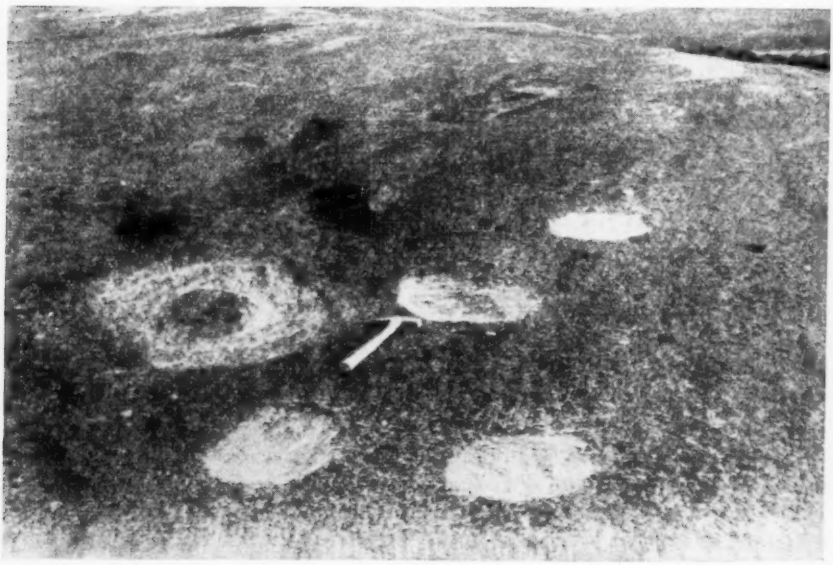

FIG. 3.—Small weather pits; same locality as Fig. 2. The surface of the granite in the pits has been scraped clean of stain.

casually by F. E. Matthes[4] and J. G. Lester.[5] Lester and Matthes ascribe them to solution, and Smith observes a relationship between exfoliation spalls and the incipient stages of the pits, but none of these writers associates them with indurated veneers. Anderson[6] describes similar depressions from the Cassia City of Rocks. He refers to them as "bathtubs" and properly associates them with indurated veneers, although he attributes their sculpture to running water.

Most of the weather pits which appear in unjointed granite masses throughout the southeastern states seem to have resulted from the breaching of indurated veneers, and it is the writer's opinion that most of this breaching has taken place through the agency of patches of moss or similar vegetation.

In the vicinity of Stone Mountain in Wilkes County, North Carolina, there are unjointed granite bosses which have many small weather pits, such as those shown in Figures 2 and 3. On the same outcrops there are many small circular patches of moss, which seem to be the agency that brings about the breaching of the veneer to make the weather pits.

If, as Anderson suggests, the veneer has been formed through the deposition of soluble mineral matter, it could readily be removed by any agency which enables the upward-moving capillary water to leave the surface of the rock in the liquid state. Because the moss usually remains somewhat moist and protects the rock surface beneath it from insolation, it is probable that the rising capillary water passes directly into it from the rock surface. The water thus enabled to pass through the veneer as a liquid should be able to remove in a short time the soluble mineral matter which previous evaporation would have taken a long time to deposit. Thus it would not seem incredible that the residence of a single moss patch should so weaken the protective veneer beneath it that, after removal of the moss, the rock would be subject to granular disintegration and therefore more vulnerable to erosion than the surrounding veneer-covered rock.

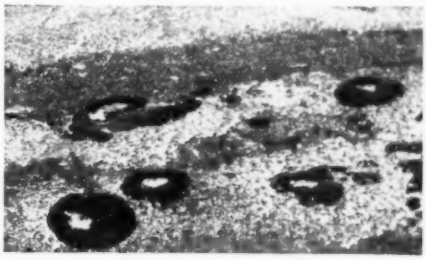

FIG. 4.—Circular moss patches on granite outcrop; same locality as Fig. 2.

On the other hand, if, as the writer believes, the veneers have been formed largely by the deposition of insoluble iron oxides, the process of their removal would be somewhat similar but probably less rapid. The ferric oxide could not be removed by the water passing into the moss until it had first been acted upon by organic acids—either the humic or carbonic acids produced by the peaty material which characteristically underlies the moss or the carbonic acid produced by the living rootlets.

Fortunately, in the Wilkes County locality there are similarities in the shapes of the weather pits and the patches of moss which support the theory relating them. Many of the pits are of the

[4] "Geologic History of the Yosemite Valley," *U.S. Geol. Surv. Prof. Paper 160* (1930), pp. 63–64.

[5] "The Geology of the Region around Stone Mountain, Georgia," *Univ. Colo. Studies*, Vol. XXVI (1938), p. 89.

[6] P. 58 of ftn. 1 (1931).

FIG. 5.—Dumbbell-shaped weather pit; same locality as Fig. 2

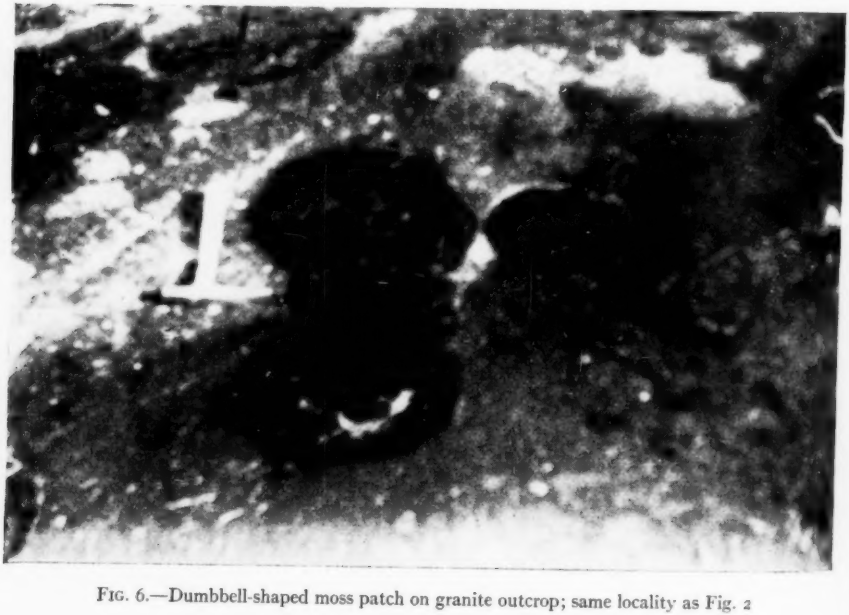

FIG. 6.—Dumbbell-shaped moss patch on granite outcrop; same locality as Fig. 2

type shown at the left in Figure 3, toward which the pick is pointed. They differ from the usual type in that they retain a patch of veneer in the center of a ring-shaped depression. Compare with these the moss patches of the type shown in Figure 4. Such ring-shaped moss patches, each with a bare spot in the center, suggest a genetic relation with the ring-shaped weather pits. The residual patches of veneer in the centers of the pits thus may have been preserved because they never were covered by moss.

Still further credence is lent to the relationship between the mosses and the pits by the fact that both occasionally occur in dumbbell-shaped forms. Such a pit is shown in Figure 5, and a similarly shaped moss patch is shown in Figure 6. Somewhat similar compound pits could, of course, be produced by partial superimposition of a young pit on an older one or by lateral growth and coalescence. However, in the example shown in Figure 5 these possibilities are minimized by the rather long, attenuated connecting passage between the two principal parts of the pit. The further possibility of a drainageway connecting these two pits is precluded by the fact that the line connecting them is normal to the slope.

FIG. 7.—Residual granite boulder showing breached veneer, near Trap Hill, North Carolina

Because weather pits retain surface water longer than the general surface of the granite, there is a tendency for moss patches to localize in the bottoms of established pits. This leads to the development of younger pits in the bottoms of older ones. Such a compound pit may be seen adjacent to the drainageway in Figure 1.

All the weather pits illustrated in this paper are little more than incipient. However, the larger pits, such as those which occur on the "flat-rocks" of the lower Piedmont, show veneers in their

overhanging rims; and the writer believes they are a more mature development of such pits as those described here.

ORIGIN OF PEDESTAL ROCKS

Figure 7 shows a result of breaching the indurated veneer on the nearly verti-

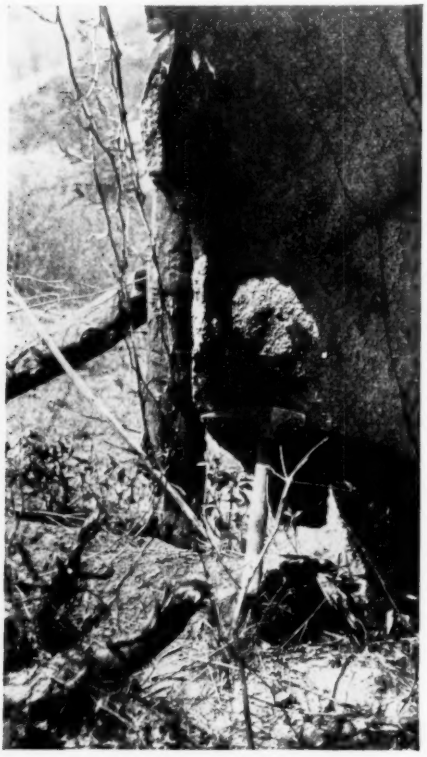

FIG. 8.—Close-up view of veneer; same exposure as Fig. 7.

cal side of a spheroidal granite outcrop. The reason for the breaching is not clear in this case; but quite possibly it was accomplished by spalling, since an exfoliation crack is visible adjacent to the opening. In other instances, where breaches appear near the bottoms of such spheroidal outcrops, they seem to have resulted from the exposure of unveneered surfaces by rapid erosion of surrounding soil. It is interesting to note in Figure 7 the tendency for the veneer to reform near the bottom of the breach at the center of the photograph. This new veneer has in turn been breached. It is not clear how the sap rock was removed from the breached area to make the deep reentrant shown; but, because of the long protective overhang of the veneer, it is presumed that the material removed to form the arched cavity must have fallen by gravity after granular disintegration.

E. Blackwelder[7] noted a relationship between hardened rock surfaces and cavernous openings in desert exposures, and J. J. Petty[8] observed an iron-stained crust on pedestal rocks of the southeastern Piedmont but does not suggest that it might have any genetic connection with the peculiarities of their form. G. W. Crickmay,[9] on the other hand, referring to Petty's observation, states that surface hardening has not been a determining factor on the form of the granite pedestal rocks of the southeast. He prefers K. Bryan's[10] explanation, which accounts for the undercutting by the differential evaporation of rainwash, which is said to be more rapid upon the cap than on the shaft, thus causing greater hydration on the latter.

In the writer's opinion the pedestal rocks are caused by the further operation of the same process which produces such breached boulders as that shown in

[7] "Cavernous Rock Surfaces of the Desert," *Amer. Jour. Sci.*, Vol. XVII (1929), pp. 393–99.

[8] "Pedestal Rocks of Granite in the Southern Piedmont," *Jour. Elisha Mitchell Sci. Soc.*, Vol. XLVIII (1932), pp. 119–22.

[9] "Granite Pedestal Rocks in the Southern Appalachian Piedmont," *Jour. Geol.*, Vol. XLIII (1935), pp. 745–58.

[10] "Pedestal Rocks in the Arid Southwest," *U.S. Geol. Sur. Bull. 760* (1925), pp. 1–11.

Figures 7 and 8. The indurated veneer can easily be seen on all the pedestal rocks he has observed, and their form seems to have resulted from the breach encircling the exposure.

Evidence for the influence of the veneer in determining the form of pedestal rocks may be found in the fact that it is usually more strongly developed on the southern side of the cap exposure than on the northern. The photograph in Figure 9, which was taken looking west, is typical. The indurated veneer which forms the protective cap extends much farther down on the southern side than on the northern, probably because of the greater evaporation on surfaces which are directly subject to insolation.

FIG. 9.—Pedestal rock in eastern Wake County, North Carolina, 2 miles southwest of Wendell

Because of the tendency for the underlying rock to disaggregate and fall away, leaving such overhanging lips of veneer as that pictured in Figure 8, it is possible that veneers have frequently been mistaken for exfoliation spalls.

THE TERMINOLOGY OF LATE-MAGMATIC AND POST-MAGMATIC PROCESSES

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ABSTRACT

Too many conflicting terms are used by petrologists and ore geologists in relation to late-magmatic and post-magmatic processes. An attempt is made to evaluate these terms and to show which are worthy of preservation and which might be eliminated.

INTRODUCTION

It is important that petrologists and ore geologists, when discussing the genesis of rocks and ores, should use the same terms in the same sense. The two terms that have been longest and most widely used by both groups are "pneumatolytic" and "hydrothermal," but these have become so warped that they mean different things to different men. Attempts to clear up these differences have generally led to the introduction of additional terms, among which are "deuteric," "pneumotectic," "autometamorphic," "late-magmatic," "epimagmatic," and "supercritical," and to the use of qualifying adjectives, such as "high-temperature hydrothermal" and "low-temperature hydrothermal." In the interest of clear thinking, a further attempt should be made to codify these terms and to eliminate some.

HYDATOTHERMAL, PNEUMATOLYTIC

The oldest term of this group is "hydatothermal." It is obsolete, but it is sometimes cited as the original form of "hydrothermal." This is correct in an etymological sense but incorrect as regards the connotation of the term. "Hydatothermal" was used tentatively by Bunsen[1] in 1847 for a group of volcanic phenomena which he afterward termed "pneumatolytic." Here is perhaps the root of a misunderstanding which has persisted ever since.

In 1851 Bunsen[2] abandoned his earlier term, replacing it by "pneumatolytic," without any change of meaning. To be sure of the meaning, one should read Bunsen's whole article. On page 239 he described the action of steam generated by hot lava coming into contact with cold, water-soaked rocks. From page 241 onward he discussed fumarole and solfatara gases and their destructive action on surrounding rocks. On page 259 he described the alteration of lava by fumarole gases, the end-product being an almost iron-free clay. On another page the formation of alunite by the action of sulphuric vapors was described.

These excerpts make it clear that the term "pneumatolytic," like "hydatothermal" before it, was applied to the destructive action of steam and acid gases upon solid rocks under atmospheric conditions, the action proceeding from without inward. Complete misunderstanding of Bunsen's meaning has been shown by certain writers—for instance, by T. Crook,[3] who has written: "To this action

[1] "Ueber den inneren Zusammenhang der pseudovulkanischen Erscheinungen Islands," *Annalen der Chemie und Pharmacie*, Vol. LXII (1847), p. 17 n.

[2] "Ueber die Processe der vulkanischen Gesteinbildung Islands," *Poggendorffs Annalen*, Vol. LXXXIII (1851), p. 238.

[3] *History of the Theory of Ore-Deposits* (London: Murby, 1933), p. 72.

of the deposition of metals from fluoriferous and other vapours the name 'pneumatolysis' was applied by Bunsen." Crook seems to confuse Bunsen with Daubrée.

In 1890 W. C. Brögger[4] found himself in need of a word to describe the class of minerals, characteristically developed in granite- and syenite-pegmatites, which contain fluorine, chlorine, boron, and other so-called "mineralizers" ("agents minéralisateurs"). For this purpose he borrowed Bunsen's term "pneumatolytic," saying: "Ich bezeichne mit diesem alten Terminus die durch die genannten *agents minéralisateurs* gebildeten Substanzen." It is not clear whether Brögger supposed the medium through which these mineralizers are conveyed to have been gaseous or liquid. But Paul Niggli[5] claims that "in all but very alkaline systems, pneumatolytic solutions may be assumed to be in the fluid state corresponding to super-critical conditions."

The differences between these three concepts of pneumatolysis are considerable. The pneumatolysis of Bunsen is characteristic of fumaroles, and it involves the destruction and leaching of magmatic silicates. The pneumatolysis of Brögger and Niggli, on the other hand, is a deep-seated phenomenon; it leads to the building-up of new crystalline silicates which may incorporate fluorine, chlorine, boric acid, phosphoric acid, or other "mineralizers." Between Brögger and Niggli the difference is that Brögger emphasized the peculiar chemical composition of the fluid, whereas Niggli stresses the supercritical state of the fluid rather than its chemical composition.

Most students of ore deposits seem to follow Brögger's definition of "pneumatolytic," and they never fail to use that adjective if they come across tourmaline or topaz in their studies; but they are divided on the question of supercritical or subcritical conditions. Experimental evidence which might decide the question is lacking. Chitarow and Ivanow[6] found that a weak solution (2.58 per cent) of potassium chloride has a critical temperature of 414° C.; but M. Benedict[7] showed that a saturated solution of the same salt does not show critical phenomena at all, the vapor pressure rising to a maximum of 255 atmospheres at 565° C. and then declining again. It cannot be doubted that the residual fluid from a feldspathic magma is rich in alkalis (whether as aluminates, silicates, hydroxides, carbonates, or chlorides); hence, it is unlikely that the critical temperature of this fluid can ever be reached. But in that case the term "pneumatolytic" is unjustified, the condition being actually high-temperature hydrothermal. The objections to "pneumatolytic" and "pneumatolysis" apply equally to the form "autopneumatolysis," used by Lacroix in 1907.

HYDROTHERMAL, HYPOTHERMAL

The adjective "hydrothermal" seems to have been used for the first time by R. Murchison,[8] in 1867. Discussing gold deposits, Murchison wrote that the metal "must have been formed into veins, geodes, and strings at a comparatively recent period, and, I think, by igneous or hydrothermal action from beneath." No

[4] "Die Mineralien der Syenitpegmatitgänge," *Zeitschr. f. Krist.*, Vol. XVI (1890), p. 159.

[5] *Ore Deposits of Magmatic Origin* (London: Murby, 1929), p. 8.

[6] "Kritische Temperatur des Wassers und der wässerigen Lösungen des kieselsauren Natriums," *Zentralb. Min.*, Abt. A (1936), p. 46.

[7] "Properties of Saturated Aqueous Solutions of Potassium Chloride," *Jour. Geol.*, Vol. XLVII (1939), p. 252.

[8] *Siluria* (4th ed.; 1867), p. 459.

definition of "hydrothermal" was offered; Murchison must have thought the literal meaning sufficient. The first attempt at a strict definition, as far as the writer knows, was made by P. Niggli and G. W. Morey,[9] who described the hydrothermal synthesis of minerals "at temperatures above 100°, up to the respective critical temperatures." Niggli has reaffirmed in later writings his contention that the boundary between pneumatolytic and hydrothermal conditions is defined by the critical temperature of the fluid. A. E. Fersmann[10] agrees with Niggli on this point and places the upper limit of the hydrothermal region, for granitic magmas, at about 400° (370°–430°).

But other writers extend the hydrothermal region upward until it meets the magmatic region. H. Ries includes tourmalinization and greisenization among hydrothermal processes. W. T. Schaller says that hydrothermal replacement reactions are essential to the development of such minerals as pink tourmaline, lepidolite, beryl, columbite, and cassiterite. Lindgren[11] divides hydrothermal deposits into "hypothermal," "mesothermal," and "epithermal," these terms indicating deposition under high temperature and high pressure, medium temperature and medium pressure, low temperature and low pressure, respectively. Among the hypothermal deposits, Lindgren includes many that were formerly considered pneumatolytic, such as cassiterite veins with tourmaline and topaz, and wolframite-scheelite deposits. Fersmann,[12] too, makes three subdivisions in the hydrothermal stage, calling them "high-thermal," "middle-thermal," and "low-thermal," respectively; but he expressly excludes from the high-thermal division all deposits of cassiterite, wolframite, and scheelite, holding that these have been formed under supercritical conditions.

ORTHOTECTIC, PNEUMOTECTIC

In 1918 Graton and McLaughlin[13] introduced the terms "orthotectic" and "pneumotectic," defining them as follows:

Orthotectic—those processes and products, strictly magmatic in the narrowest sense, exemplified in the normal crystallization of normal igneous rocks like granite and diorite.

Pneumotectic—those next-subsequent processes and products of magma consolidation in which fundamental influences of a sort that was magmatic in the strictest sense were recognizably modified and to some extent controlled by gaseous constituents or so-called mineralizers.

LATE-MAGMATIC

Following the last definition, the pneumotectic stage is a late-magmatic stage. Unfortunately, the term "late-magmatic" was used in another sense by C. F. Tolman and A. F. Rogers,[14] who wrote that "the magmatic ores in general have been introduced at a late magmatic stage as a result of mineralizers, and the ore minerals replace the silicates." If the ores replace the silicates, then the process is post-magmatic rather than late-magmatic; and, since the co-operation of mineralizers is involved, the process described by Tolman and Rogers seems to be indistinguishable from the pneuma-

[9] "Die hydrothermale Silikatbildung," *Zeitschr. f. anorg. Chem.*, Vol. LXXXIII (1913), p. 369.

[10] "Über die geochemisch-genetische Klassifikation der Granitpegmatite," *Min. Pet. Mitt.*, Vol. XLI (1931), p. 64.

[11] *Mineral Deposits* (New York: McGraw-Hill Book Co.), 3d ed. (1927), p. 135; 4th ed. (1933), p. 210.

[12] P. 64 of ftn. 10 (1931).

[13] "Further Remarks on the Ores of Engels, California," *Econ. Geol.*, Vol. XIII (1918), p. 85.

[14] *Magmatic Sulfid Ores* (Stanford University: Stanford University Press, 1916), p. 7.

tolysis of Brögger. It falls also within the definition of deuteric and of high-temperature hydrothermal.

EPIMAGMATIC

Fersmann[15] has applied the term "epimagmatic" to that stage of magmatic crystallization which (in a granitic magma) immediately precedes and in part includes the generation of aplite and graphic granite. He puts the temperature limits of this stage between 800° and 700°. Since at this late stage of crystallization the concentration of fugitive components in the magma must be at its maximum, the connotation of epimagmatic would appear to be almost identical with that of pneumotectic.

DEUTERIC, ENDOMETAMORPHIC

The term "deuteric" was introduced in 1916 by J. J. Sederholm,[16] who defined it to mean "changes which have taken place in direct continuation of the consolidation of the magma itself." This definition clearly covers both the pneumatolytic and the hydrothermal stages of other writers, including the late-magmatic stage of Tolman and Rogers. It differs from "pneumatolytic" (Brögger) by not insisting on the participation of mineralizers other than water; and from "pneumatolytic" (Niggli) by ignoring the distinction between supercritical and subcritical conditions. It does not differ from "hydrothermal" as used by Ries and by Schaller. But "deuteric" is a petrologist's term; it applies specifically to the internal reactions that affect magmatic rocks, whereas "hydrothermal" is applied indifferently to rocks and to ore deposits, whether the latter occur within or without the intrusive rock.

R. J. Colony[17] proposed to extend the term "deuteric" to "all magmatic end-stage phenomena," including even contact metasomatism and granitization. Sederholm was disinclined to accept this extension of his term; yet it must be admitted that deuteric reactions such as tourmalinization, taking place within a body of granite, are often associated with metasomatism and mineralization of the rocks in the contact zone. The terms "endomorphism" (properly "endometamorphism") and "exomorphism" (properly "exometamorphism") are sometimes used to distinguish internal from external reactions of this nature.

AUTOMETAMORPHIC, ETC.

Still another term, "autometamorphism," has been applied to the spilitic alteration of basic lavas by their own residual solutions (H. C. Sargent[18]). If this term is interpreted literally, it has the same meaning as "deuteric"; but the spilitic alteration of basic lavas is a low-temperature reaction characterized by the formation of chlorite and albite. Used in this sense, "autometamorphic" has a much narrower connotation than "deuteric." But C. S. Ross[19] finds a use for both terms—one for reaction in a closed system (i.e., without introduction of extraneous matter) and the other for reaction in an open system.

To complete this list of terms, we may add "hypogene" and "supergene," applied by Ransome[20] to changes taking

[15] Ftn. 10 (1931).

[16] "On Synantetic Minerals," *Geol. Comm. of Finland, Bull. 48* (1915), p. 141.

[17] "The Final Consolidation Phenomena in the Crystallization of Igneous Rocks," *Jour. Geol.*, Vol. XXXI (1923), p. 169.

[18] "On a Spilitic Facies of Lower Carboniferous Lava-Flows," *Quart. Jour. Geol. Soc. London*, Vol. LXXIII (1918), p. 19.

[19] Personal communication.

[20] "Copper Deposits near Superior, Arizona," *U.S. Geol. Soc. Bull. 540* (1912), p. 152.

place under the influence of ascending and descending waters, respectively; also "hypergene," used by Fersmann in the same sense as "supergene."

DEFINITION BY TEMPERATURE

The number of terms already enumerated seems excessive in relation to the number of clearly recognizable stages in the cooling of eruptive magma and the generation of eruptive rocks and ore deposits. If it were practicable to calibrate the geological thermometer scale at frequent intervals from 700° downward, then it might be useful to have a term corresponding to each interval; but the only fixed points in this part of the scale are the high-low quartz transformation (573°) and the critical temperature of pure water (374°). The first is somewhat affected by pressure, and the second is greatly affected by the presence of dissolved salts. Nevertheless, Fersmann has set up an arbitrary scale of temperature from 800° downward and distinguishes eleven phases in the crystallization of granite-pegmatite, as below:

		Temperature
Magmatic solutions	Phase A (magmatic)	Above 800°
	Phase B (epimagmatic)	800°–700°
	Phase C (pegmatitic)	700°–600°
Fluid solutions	Phase D–E (pegmatoid)	600°–500°
	Phase F–G (supercritical)	500°–400°
Aqueous solutions	Phase H (high thermal)	400°–300°
	Phase I (middle thermal)	300°–200°
	Phase K (low thermal)	200°–100°
	Phase L (hypergene)	Below 100°

One would gladly follow the lead of such a distinguished authority as Fersmann if these phases could be separated in practice as easily as on paper. Phases B and C are distinguishable by texture, and C and D from E by high or low quartz. But how is one to distinguish *in practice* between E, F, G, and H? Fersmann has named certain minerals which he considers typical of each phase:

Phase C—minerals containing Ti, Nb, Ta, Y, rare earths
Phase D—schorl, mica
Phase E—topaz, beryl
Phase F—lithium compounds
Phase G—lithium-manganese phosphates
Phase H—cryolite, fluocarbonates
Phase I—sulphides of Pb, Cu, Zn
Phase K—zeolites

ORDER OF CRYSTALLIZATION IN PEGMATITES

It is a matter of fairly general agreement that the lithium silicates follow black tourmaline (schorl) and that these silicates are followed by phosphates, sulphides, and carbonates; but there is not complete agreement about the details of this succession. For instance, W. C. Brögger put the zeolite phase (K) before the fluocarbonate phase (H) in the Norwegian syenite-pegmatites; K. K. Landes places columbite (C) after schorl and beryl (D and E) in the pegmatites of Maine; and T. W. Gevers puts beryl before tantalite and rare-earth minerals in Namaqualand. R. H. Jahns has found that microlite and tantalite (C) follow spodumene (F) in the Harding pegmatite, New Mexico. W. T. Schaller[21] affirms that "any one of these [pegmatite] minerals may have more than one

[21] "Mineral Replacements in Pegmatites," *Amer. Min.*, Vol. XII (1927), p. 62.

generation, and reversals of order of formation also occur." Even graphic granite, which Fersmann considers to have been formed above 700° (phase B), has sometimes been generated by hydrothermal replacement of aplite dikes, as in a remarkable instance recently described by N. M. Uspensky.[22] As for the suggested temperature limits, few people are likely to agree that schorl (black tourmaline) always indicates a temperature of 500°–600°. A. Brammall and H. F. Harwood[23] showed that in the Dartmoor granite tourmaline occurs under the following conditions: (*a*) as an apparently primary mineral, (*b*) in geodes, (*c*) in tourmalinized granite, (*d*) in quartz-tourmaline veins, and (*e*) in country rocks. These observations point to a wide range of temperature. In the Black Hills of South Dakota, V. Ziegler[24] showed that tantalite, tourmaline, muscovite, spodumene, and lepidolite all occur in quartz veins as well as in pegmatites. In the Enoggera granite of Queensland, black tourmaline occurs in vugs along with calcite, pyrite, and zeolites; it was formed "after pyrite and before calcite" (M. Whitehouse[25]). A still later period of crystallization is shown by the overgrowths of new tourmaline upon detrital grains, which P. D. Krynine[26] has described in Paleozoic sandstones of Pennsylvania.

[22] "On the Genesis of Granite Pegmatites," *Amer. Min.*, Vol. XXVIII (1943), p. 445.

[23] "On the Temperature Range of Formation of Tourmaline in Dartmoor Granite," *Min. Mag.*, Vol. XXI (1927), p. 205.

[24] "The Differentiation of a Granitic Magma," *Jour. Geol.*, Vol. IX (1914), p. 264.

[25] "The Deuteric Mineral Sequence in the Enoggera Granite," *Min. Mag.*, Vol. XXIV (1937), p. 538.

[26] "Petrology and Genesis of the Third Bradford Sand," *Pa. State College, Bull. 29* (1940), p. 30.

The plain conclusion to be drawn from such evidence as this is that the typical pegmatite minerals are not restricted to narrow temperature limits but that each of them is stable through a wide range of temperature and may be formed anywhere within that range. A mineral which contains an uncommon component such as beryllium, boron, or lithium must owe its formation less to a particular physical condition than to a particular chemical condition which allowed that rare component to be concentrated in sufficient quantity to generate beryl, tourmaline, or spodumene. In short, the use of these relatively uncommon minerals as temperature-indicators in pegmatite is thought to be unsound in principle and misleading in practice. (Parenthetically, there seems to be a wonderful opportunity for the revival of Sorby's "contracting-bubble" method of determining the temperature of formation of pegmatite minerals. Topaz and beryl, in particular, are often crowded with fluid enclosures.)

CRYSTALLIZATION OF A HYDROUS MAGMA

The idea that the minerals of granite-pegmatites must necessarily have been formed in a constant order seems to be an inheritance from the belief that the change from magmatic conditions to hydrothermal conditions is gradual and continuous. This belief has been widely held for many years, and it is embodied in the well-known temperature-concentration diagram of Niggli,[27] which is still reproduced in some textbooks. But Niggli was careful to point out that the diagram is not applicable if unmixing takes place in the fluid. In the case of a natural water-bearing magma, unmixing ought

[27] Fig. 1, p. 3, of ftn. 5 (1929).

to be expected. R. L. Goranson's investigation[28] of the solubility of water in molten granite showed that the maximum amount of water that can be held in solution by granitic magma, under any geologically conceivable conditions, is about 9 per cent. When, in the course of crystallization of a granitic magma, the content of water reaches this maximum, then further crystallization must lead to one or another of the following consequences:

a) The excess of water may escape by diffusion into the walls of the magma chamber. If this is impossible, then—

b) The excess of water may enter the solid phase or phases as water of constitution, generating hydroxyl-bearing silicates of low water content, such as hornblende, mica, and tourmaline. If the excess is continuously used up in this way, then there will be only one fluid phase until crystallization is complete.

c) If water is not completely used up, as in (*b*), then it must separate as a second fluid phase, forming an emulsion with the magma. Under low pressure this would be a vapor phase; but under a pressure high enough to prevent boiling, the new phase, being saturated with alkali salts, would in all probability be a liquid below its critical temperature. Such an emulsion of two liquids is an interesting possibility, but it does not seem to necessitate any interruption in the crystallization of the system or any change in the character of the solids produced. Hydroxyl-bearing silicates would continue to crystallize, as before, until the magma fraction of the emulsion was used up and only the aqueous-alkaline fraction remained.

When the last of the magma has crystallized under condition (*c*), the pores of the newly formed rock will be filled with the aqueous-alkaline liquid. With further cooling, reaction must set in between the crystals and the liquid, causing such changes as albitization of feldspar; conversion of pyroxene into soda-pyroxene, amphibole, or mica; tourmalinization; etc. If uncommon elements are sufficiently concentrated in the liquid, uncommon minerals—such as beryl, spodumene, and tantalite—will crystallize at this time, especially in pegmatite veins and druses. Certain oxides and sulphides—particularly magnetite, spinel, cassiterite, pyrite, pyrrhotite, and molybdenite—may also make their appearance. This is the deuteric stage of the petrologists, corresponding to the high-temperature hydrothermal stage among ore deposits.

With still further cooling, a final stage of low-temperature hydration and leaching is reached, in which all the earlier silicates tend to break down and to be replaced by scaly, fibrous, and colloform silicates with a relatively high content of water (chlorite, 10–15 per cent; serpentine, 13 per cent; kaolin, 14 per cent; zeolites, 10–20 per cent). Carbonates are also characteristic of this stage, which corresponds to the low-temperature hydrothermal stage in ore deposition.

ARGUMENT

Considered in this light, it seems that there are only four clearly distinguishable stages in the passage of a hydrous magma into a body of rock at atmospheric (or near atmospheric) temperatures, as follows:

1. A stage when only anhydrous silicates (and possibly other anhydrous compounds) crystallize out of the magma
2. A stage when hydroxyl-bearing silicates of low water content (mostly less than 5 per cent) crystallize out of the magma in addition to anhydrous silicates
3. A stage when the same hydroxyl-bearing silicates grow in the solid rock by replacement of earlier silicates
4. A final stage when all the previously formed silicates become unstable and are replaced by scaly, fibrous, or colloform decomposition products of higher water content (more than 10 per cent) and by carbonates

[28] "The Solubility of Water in Granite Magmas," *Amer. Jour. Sci.*, Vol. XXII (1931), p. 481.

The first and second stages are both magmatic ("orthomagmatic," Niggli). The first stage has never received a distinctive name, even Graton and McLaughlin's term "orthotectic" being so defined that it covers the second, as well as the first, stage. The second stage conforms closely to the epimagmatic stage of Fersmann and the pneumotectic stage of Graton and McLaughlin but cannot be identified completely with either of these. It also satisfies the definition of "orthotectic."

The third stage is the one which carries the greatest number of labels. "Pneumatolytic," "autopneumatolytic," "supercritical," "pegmatitic," "late-magmatic," "deuteric," "endo(meta)morphic," and "hypothermal" or "high-temperature hydrothermal" all apply to this stage. The objections to "pneumatolytic" have already been pointed out; the term has been used in so many senses that its value has been destroyed and it has become a menace to clear thinking. "Supercritical" should not be used in the absence of proof (or at least a high probability) that this condition was actually attained. "Pegmatitic" is useful as a textural term; but it is not strictly applicable to a stage of crystallization, since there is reason to think that pegmatite may be formed in more than one stage. "Late-magmatic" is open to objection, since the stage is actually post-magmatic. "Deuteric" is unobjectionable from the petrologic viewpoint but is not applicable to deposits formed or processes operating outside of an igneous body. "Endo(meta)morphic" is synonymous with "deuteric." "Autometamorphic" in its literal sense may be applied to this stage, but in the narrower sense of spilitic alteration it refers to the fourth stage. "Hypothermal" combines the two concepts of temperature and depth; consequently, it may be applied to a particular deposit but not to a particular process or a particular stage of crystallization. "High-temperature hydrothermal" avoids this objection and is applicable either to a process, a stage, or a deposit. The only disadvantage of this term is its length. But ore geologists will not readily give up "hypothermal," and there seems no reason why petrologists should not retain "deuteric" for use in their special province.

The fourth stage, which is a stage of degradation and death for rocks but one of birth for many ore deposits, is also hydrothermal. It should be distinguished as "low-temperature hydrothermal," or more fully as "intermediate and low-temperature hydrothermal"; but it is questionable whether a satisfactory mineralogical distinction has been established between intermediate and low-temperature conditions. Autometamorphism, as described by Sargent, is a low-temperature hydrothermal process.

It does not seem possible to define these stages by precise temperature limits. A magma with a small content of water will reach the second stage later than one with a large content of water, or it may pass directly from the first to the third stage, omitting the second. Similarly, the deuteric or third stage may be more prolonged in one rock than in another, or it may be suppressed altogether. For example, in the Cortlandt complex at Peekskill, New York, the simple pyroxene norite completed its crystallization in the first stage; the dioritic rocks (those with primary hornblende) passed into the second stage but suffered little third-stage alteration; whereas the rocks with poikilitic hornblende seem to have omitted the second stage and to have passed directly from the first to the third.

CONCLUSIONS

The conclusions to which the writer has been led may be summarized as follows:

1. That the "pneumo-" terms—"pneumatolytic," "autopneumatolytic," and "pneumotectic"—should be finally abandoned, on the ground that they express an assumption which has never been proved;

2. That "orthotectic" and "epimagmatic" are unnecessary; and that "autometamorphic" and "late-magmatic" are not clearly distinguished from "deuteric" and should be abandoned in favor of the latter term;

3. That terms which combine the two concepts of temperature and pressure, such as "hypothermal," "mesothermal," and "epithermal," should be reserved for types of ore deposits. They are too indefinite for petrologic or physicochemical use;

4. That the four stages of crystallization and reaction that were defined above (p. 348) might be indicated in a simple way by the following terms and symbols: "first magmatic stage" (M_1); "second magmatic stage" (M_2); "deuteric or high-temperature hydrothermal stage" (H_1); and "low-temperature hydrothermal stage" (H_2).

GEOLOGIC SIGNIFICANCE OF A COARSE MARINE SEDIMENT FROM NEAR SANTA CATALINA ISLAND, CALIFORNIA[1]

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ABSTRACT

Dredge samples of a coarse sediment taken from a depth of 150 fathoms of water off the southeast end of Santa Catalina Island, California, exhibit characteristics that suggest the sediment to be of beach origin. The conclusion is reached that the island has subsided approximately 600 feet in post–Middle Miocene time, and possibly in post-Pleistocene time.

INTRODUCTION

A considerable amount of work has been done on the sediments of the sea bottom off the southern California coast by Parker D. Trask,[2] by R. Revelle and F. P. Shepard,[3] and by others.[4] It is the purpose of the present paper to discuss in some detail a sediment that hitherto has received only passing mention and yet is of considerable significance in connection with the relatively recent geologic history of the channel region.

Research work in sedimentation at the University of Southern California was started in the fall of 1940, with the establishment of a sedimentation laboratory in the new Hancock Building under the Allan Hancock Foundation for Scientific Research. At the same time, the taking of bottom samples for sedimentation studies was added to the regular collecting program of the "Velero III," the motor ship belonging to the Foundation. These included portions of the dredge samples taken in the routine hauls for specimens of marine life, as well as cores obtained with a modified Trask coring tube, patterned after that in use at the Scripps Institution of Oceanography at La Jolla. The war caused a cessation of collecting activities, with the taking over of the "Velero III" by the United States Navy.

LOCATION

The sediment herein discussed came from a submarine bank extending to the southeast from Santa Catalina Island and at a distance of approximately 6 miles from the southeastern end of the island. The general position is 33°15′ north latitude and 118°12′ west longitude (see Fig. 1). The "Velero" station number is 1188–40.

Two consecutive dredge hauls were made at this locality, the first starting at 145 fathoms and ending at 150 and the second starting at 150 fathoms and ending at 155. Similar types of sediment were taken on both hauls, although on the second the dredge was damaged and only a portion of the haul recovered. It is reported by members of the scientific staff of the "Velero III" that previous hauls in the same general locality had brought up sediments of the same type, and hauls taken at a later date likewise were similar.

[1] A contribution of the Sedimentation Laboratory, Allan Hancock Foundation for Scientific Research, University of Southern California

[2] "Sedimentation in the Channel Islands Region," *Econ. Geol.*, Vol. XXVI (1931), pp. 24–43.

[3] "Sediments off the California Coast," in *Recent Marine Sediments* (Tulsa: Amer. Assoc. Pet. Geol., 1939), pp. 245–82.

[4] F. P. Shepard and W. F. Wrath, "Marine Sediments around Catalina Island," *Jour. Sed. Pet.*, Vol. VII (1937), pp. 41–50; Shepard and G. A. McDonald, "Sediments of Santa Monica Bay, California," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXII (1938), pp. 201–16.

CHARACTER OF SEDIMENTS

The material from both hauls was split by hand in order to facilitate handling, care being taken to obtain the proper relative proportions of large and small, angular and rounded fragments. The reduced sample was taken to the laboratory for roundness, shape, and frequency

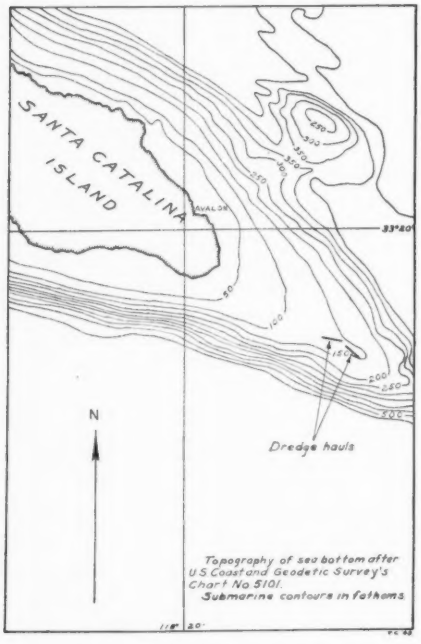

FIG. 1.—Map showing where dredge hauls of coarse sediment were made.

determinations, for hand-lens determination of the rock types represented, and for study of the Foraminifera present in some of the specimens.

The sediments fall into two grades of the Wentworth scale: pebbles (4–64 mm.) and cobbles (64–256 mm.). Because of the very considerable weight of some of the specimens, a number frequency, rather than a weight frequency, was determined. This gave approximately 70 per cent in the pebble grade and 30 per cent in the cobble grade. If finer sediments were present, they were washed out of the dredge on the way to the surface. However, indications are that no fine material exists at this locality, since none was obtained in the mud sack in any of the hauls made, then or subsequently.

Measurements made on the specimens of three axes at right angles to each other gave a range of from 6.2 mm. to 225.9 mm. The average for the long axes was 57.5 mm.; for the intermediate, 39.3 mm.; and for the short, 25.4 mm. Thus, the average specimen had one long and two short axes, or a definite ellipsoidal shape.

Roundness studies made on a representative number of the specimens, using Wentworth's roundness factor,[5] gave a range from 0.271 to 0.981, with an average of 0.742. Since a perfect sphere would have a roundness factor of 1.00, the approach to roundness of the material is evident. It should be noted, however, that these roundness factors were taken in the plane of the intermediate and longest axes; if they had been taken in the plane of the longest and shortest axes, smaller values for the roundness factor would have resulted. A classification based on the less accurate, but somewhat more easily visualized terms—angular, subangular, subrounded, and rounded—gave: 16 per cent angular, 20 per cent subangular, 40 per cent subrounded, and 24 per cent rounded. This likewise indicates the approach to roundness of the material.

The composition of a representative number of the cobbles and pebbles was determined by means of the hand lens on a freshly broken surface. The results were as follows: 12 per cent sedimentary,

[5] W. C. Krumbein and F. J. Pettijohn, *Manual of Sedimentary Petrography* (New York: D. Appleton-Century Co., 1938), p. 279.

4 per cent doubtfully metamorphic, and 84 per cent igneous.

Of the sedimentary rocks, one, the largest and most angular obtained in either haul, was a ferruginous, quartz sandstone, which presumably had not been subjected to the wear of most of the others, since it was definitely softer than many of the rounded specimens. However, it showed no fresh surface to indicate that it had been broken from a submarine outcrop by the dredge. Some of the others were interesting because of the Foraminifera they contained. These were studied by Mr. Elmer S. Franklin, who lists some forty species, calling attention to the lack of arenaceous forms and the abundance of specimens belonging to the genera *Elphidium*, *Cassidulina*, and *Quinqueloculina*. He reports that the fauna is equivalent to the Relizian of Kleinpell (Middle Miocene).[6]

The metamorphic rocks consist of some questionable schistose quartzites, the exact nature of which would be more apparent under the microscope. No typical Franciscan metamorphics were observed.

The igneous rocks fall principally into three general groups: (1) andesites and dacites, which comprise 54 per cent of the material as a whole; (2) basalts, comprising 20 per cent; and (3) diorite, making up 8 per cent. A very small amount of tuff was present also—perhaps 2 per cent. In the first group the andesite predominates over the dacite in the ratio of 7:2.

Some of the andesites and dacites are very fine grained; others are porphyritic, with phenocrysts of surprisingly fresh plagioclase. The quartz is rather sparingly present in the dacites and suggests that the latter originated as a differentiate of the same magma that gave rise to the andesites. Some of the basalt pebbles are angular and appear quite fresh; some are subrounded or rounded and are very similar in appearance to the Topango (Middle Miocene) basalts of the Santa Monica Mountains. The diorite exhibits prominent elongated hornblende crystals and some tendency to a gneissic structure.

W. S. T. Smith reported in 1933 that the southeastern part of Santa Catalina Island consisted chiefly of diorite and of andesitic lava flows.[7] The great abundance of the latter rock type and the presence of some diorite in the sediment here under discussion suggest direct derivation of the material from the island.

CONCLUSIONS

W. H. Twenhofel has stated that ellipsoidal pebbles and cobbles are characteristic of some beaches, although not necessarily typical of the beach environment.[8] If the principal force of the waves comes, in general, from one direction, approximately at right angles to the beach, then fragments of homogeneous rock rolled back and forth by the waves should tend to develop an ellipsoidal shape, with the long axis horizontal and parallel with the shoreline and the short axis in the direction of application of the principal force.

River wear, on the other hand, according to Wentworth's experiments,[9] tends to produce spheroidal, rather than ellipsoidal, shapes from homogeneous rock fragments; and the andesites and dacites are relatively homogeneous. Furthermore, the near-by source of most

[6] Personal communication.

[7] "Marine Terraces on Santa Catalina Island," *Amer. Jour. Sci.*, Vol. XXV (1933), p. 131.

[8] *Treatise on Sedimentation* (2d ed.; Baltimore: Williams & Wilkins Co., 1932), pp. 204–5.

[9] C. K. Wentworth, "Laboratory and Field Study of Cobble Abrasion," *Jour. Geol.*, Vol. XXVII (1919), pp. 507–21.

of the material involved precludes long transportation by streams, and the size of most of the fragments makes transportation by marine currents most unlikely.

The logical conclusion, then, on the basis of the evidence cited, is that the sediment represents a beach deposit. It should be pointed out, however, that, even if it were a river gravel, its geologic significance, as here interpreted, would not be changed. The presence of a beach deposit, or a river deposit, at a depth of 900 feet below sea-level can mean only a change of sea-level of this magnitude, at least locally, since the formation of the deposit.

Such a change of sea-level could have been brought about by sinking of the island as the result of crustal disturbance, by rise of the level of the sea through return of water from the melting Pleistocene ice sheets, or by a combination of the two. Both Davis and Shepard have suggested a rise of sea-level of as much as 300 feet since the last glacial epoch.[10] If this is accepted as a reasonable figure, then the remaining 600 feet must be the result of crustal disturbance.

The presence of fragments containing a Middle Miocene foraminiferal fauna indicates that the sediment is not older than Middle Miocene, and therefore the above-mentioned crustal disturbance is definitely post–Middle Miocene. On the other hand, the unconsolidated nature of the sediment, its apparent lack of any covering of later sediments, and its occurrence on what might be interpreted topographically as a submerged terrace suggest that it is much younger than Middle Miocene, although it is recognized that degree of consolidation is not a reliable criterion of age.

On the mainland, along the seaward margin of the Los Angeles Basin, Lower or Middle Pleistocene (Lower San Pedro) is represented by several hundred feet of unconsolidated or poorly consolidated sands and gravels, folded as the result of the Pasadenan orogeny and overlain unconformably by the thinner unconsolidated sands and gravels, both marine and nonmarine, of Upper Pleistocene (Upper San Pedro) age.[11] The latter, in turn, have been faulted and tilted, and in some places folded, by post-Pleistocene disturbances.[12] It is the opinion of the senior author that the sediment here under discussion may well be the correlative of the Upper San Pedro of the mainland, and the crustal disturbance, therefore, post-Pleistocene. This is not out of accord with the views of Lawson and of Reed, as expressed by the latter in his *Geology of California*.[13]

It is concluded, then, that Santa Catalina Island is part of a subsiding block, the subsidence of which, definitely in post–Middle Miocene time, and possibly in post-Pleistocene time, has amounted to 600 feet.

ACKNOWLEDGMENTS.—The writers wish to acknowledge their indebtedness to Captain G. Allan Hancock for making possible the collecting and study of the bottom samples. Acknowledgment is made also of the untiring help given by members of the scientific staff and of the crew of the "Velero III." Mr. Elmer S. Franklin, graduate student at the University of Southern California, aided in the interpretation of the sediments through his work in identifying the Foraminifera present in some of the specimens.

[10] W. M. Davis, "Glacial Epochs of the Santa Monica Mountains," *Bull. Geol. Soc. Amer.*, Vol. XLIV (1933), p. 1045; Shepard and Wrath, p. 44 of ftn. 4 (1937).

[11] R. D. Reed and J.S. Hollister, *Structural Evolution of Southern California* (Tulsa: Amer. Assoc. Pet. Geol., 1936), p. 135.

[12] *Ibid.*, p. 50.

[13] (Tulsa: Amer. Assoc. Pet. Geol., 1933), pp. 3–8.

GLACIAL EPOCHS OF THE SANTA MONICA MOUNTAINS
A DISCUSSION

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ABSTRACT

The emergence of the coastal features and ranges of southern California and the events of their subsequent history cannot be fitted into postglacial time. It is suggested that the Coast Range orogeny took place in the Yarmouth age.

The Pasadenan orogeny, of Hans Stille,[1] or "Coast Range orogeny," as it has been termed more recently by T. L. Bailey,[2] is known to have occurred in middle or even perhaps rather late Pleistocene times; for some 5,000 feet of definitely Pleistocene marine strata in the Ventura basin (which follow the Pliocene conformably) have been shown by W. C. Putnam[3] and by Bailey[4] to be involved in the folding that has accompanied the uplift of the Coast Ranges of southern California. As R. D. Reed and J. S. Hollister[5] have noted when discussing the strong unconformity within the marine Pleistocene already known at San Pedro: "In most other California localities where Pliocene beds are folded the conditions can all be satisfactorily explained on the assumption that the last strong folding of the Cenozoic was contemporary with the folding at San Pedro."

In the Ventura district Putnam has shown also that warping as well as uplift has affected recently cut river terraces[6] and marine terraces,[7] on some of which the elevation changes at the rate of 100 feet per mile—tilting comparable to that measured near Wellington, New Zealand, on another shoreline recently affected by a renewal of "transverse deformation."[8] Earthquake phenomena in addition confirm the view that there has recently been a recurrence of orogenic movement in the region. This has taken place after a prolonged pause, however, as Putnam has shown in his account of the Ventura geomorphology. So it would be wrong to assume that the Coast Range orogeny is still in progress. Like older paroxysms of folding, it takes its place among the past events of geological history.

Bailey's paleontological evidence has recently established the intra-Pleistocene (as opposed to a hypothetical pre-Pleistocene) age of the large-scale folding movements; and he shows, moreover, that the steeply tilted Pleistocene beds at the top of the Santa Barbara

[1] "Der derzeitige tektonische Zustand der Erde," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XX (1936), pp. 849–80.

[2] "Late Pleistocene Coast Range Orogenesis in Southern California," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 1549–68.

[3] "Geomorphology of the Ventura Region, California," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), pp. 691–754.

[4] Ftn. 2 (1943).

[5] *Structural Evolution of Southern California* (Tulsa: American Association of Petroleum Geologists, 1936), p. 49.

[6] Ftn. 3 (1942).

[7] Putnam, "The Marine Cycle of Erosion for a Steeply Sloping Shoreline of Emergence," *Jour. Geol.*, Vol. XLV (1937), pp. 844–50.

[8] C. A. Cotton, "Shorelines of Transverse Deformation," *Jour. Geomorph.*, Vol. V (1942), pp. 45–58.

formation at Ventura contain a cold-water fauna which may be interpreted as of glacial age.

In this section at Ventura, Bailey has found only one such horizon, however; and it is not to be assumed without question that the orogeny has been entirely postglacial. If it were so, the title given by W. M. Davis to his study of the marine terraces of the adjacent coast would be not only facetious but quite fallacious. He named this study "Glacial Epochs of the Santa Monica Mountains"[9] and confidently assumed that individual terraces, the cutting of which he attributed to episodes of rising sea-level, might be correlated with world-wide cycles of deglaciation. The foundation of this theory must collapse, of course, if it be true that the Santa Monica Mountains had not risen and that the initial forms of the shoreline of southern California had not emerged until after the glacial period.

It seems unnecessary, however, if we may judge from published information, to assume that the now folded marine Pleistocene of southern California represents the results of deposition throughout the whole of the glacial period. The one horizon (or two, according to some authorities) of cold-water fauna identified in it may very well date from pre-Yarmouth stages of glaciation. It seems quite probable that the long Yarmouth interglacial age included the period of accumulation of the San Pedro formation (as defined by Bailey), and it perhaps also witnessed the emergence of much of southern California from the sea and the reduction of its infantile ranges to mature forms of relief.

According to current theories, there have been sufficiently great climatic oscillations in post-Yarmouth times to admit at least the possibility—perhaps even the plausibility—of Davis' theory that eustatic rise of ocean level has several times resulted in submergence at the shoreline in this region in spite of a continuous uplift of the land.

Whether positive movement of sea-level is the full, necessary, and only true explanation of episodes of shoreline retrogradation is another matter, however. In view of the fact that "approximately thirteen marine terraces of post-folding age are found along the Southern California coast," Reed and Hollister[10] "reflect on the extreme difficulty of correlating any of them with definite stages of the Ice Age."

The only argument that can be adduced in favor of relegating the Coast Range orogenic uplift with its long train of erosional consequences, as well as all the later history written in marine terraces, to the extremely short interval that has elapsed since the recession of the northern continental glaciers is the apparent absence of faunal evidence of cold climates in unconformable Upper Pleistocene accumulations.[11] It may be pointed out in this connection, however, that if there is truth in the assumption made by Davis, the beds containing relics of faunas preserved on terraces—the only traces of late Pleistocene marine life that could be preserved—must have accumulated in the warm episodes of deglaciation to which terrace-cutting has been confined.

[9] *Bull. Geol. Soc. Amer.*, Vol. XLIV (1933), pp. 1041–1133.

[10] P. 50 of ftn. 5 (1936).

[11] A "post-Pleistocene" age has been suggested for the molluscan fossils on the marine terraces at Ventura by U. S. Grant, in Putnam, p. 700 of ftn. 3 (1942).

PLEISTOCENE COAST RANGE OROGENESIS IN SOUTHERN CALIFORNIA: A DISCUSSION

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The contributions which have led to the preceding discussion by Professor Cotton call forth a further comment. The discovery that notable mid-Pleistocene folding took place in certain parts of southern California, where almost uninterrupted sedimentation had prevailed from the Pliocene far into Pleistocene time, was surprising to not a few geologists. Holding the general view that the close of the Tertiary period was characterized rather widely by an episode of orogenesis near the end of the Pliocene epoch, they thought it strange that the southern California Coast Range should instead display folding in Middle (or even later) Pleistocene time. Accumulating evidence, however, has dated this folding satisfactorily. To this evidence, T. L. Bailey has recently added important new facts which are very welcome; but the concluding portion of the abstract of his valuable paper[1] raises questions which seem to require further consideration. Quoting from that abstract:

> Paleontologic, stratigraphic, and structural evidence clearly indicate that the principal orogeny in the transverse (east-west) Coast Ranges of southern California occurred during the Pleistocene, not at the beginning as has been commonly believed. This evidence suggests that two of our widely held geological concepts need to be modified.
>
> (1) Extensive uplift and erosion may take place within the geologic lifetime of one vertebrate species in tectonically active areas such as California.
>
> (2) A great angular unconformity with differences in dip as high as 60° may occur within a single geologic period instead of only at the beginning or end of a period, as is commonly assumed.

The recency of the Pleistocene and consequent fulness of its interpreted history have led to an exaggerated idea of its importance in the long history of the earth. Quite commonly the "Glacial *period*" has been treated as if it were a full-fledged geologic period like the Cambrian, Silurian, or Cretaceous. But this is by no means the case, if we think to scale. About thirty million years has been the duration of the shorter geologic periods, while some of the longer ones have covered twice that span. Most periods are divided into three or four epochs. Ten million years would thus be about the average length of an epoch. The Pleistocene's one million years would not go far in making even an ordinary epoch on this basis. What we now call the Pleistocene should seemingly be considered only a *part* of a geologic *epoch*.[2] Perhaps several million years more will have to elapse before it becomes finally evident whether what we know as the Pleistocene is best considered (1) a finishing portion of the Pliocene epoch; (2) the early part of a separate epoch of high-standing lands completing the Tertiary period; or (3) the beginning of an epoch in a new post-Tertiary (Quaternary or otherwise-named) period which has only just started. To whichever period it more properly be-

[1] "Late Pleistocene Coast Range Orogenesis in Southern California," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 1549–68.

[2] See review of *Annual Reports of the Iowa Geological Survey, 1934–1939, with Accompanying Papers*, by R. T. C., *Jour. of Geol.*, Vol. L (1942), pp. 918–19.

longs, it was a time of stress and storm in a long transition such as has commonly occurred between geologic periods.

Orogenic episodes which aid in marking off geologic periods are commonly not simple, single events. For example, a beginning of the Taconic orogeny, which played its part in closing the Ordovician period, was evidenced by the resulting sediments as early as the start of the Cincinnatian epoch. A notable pulsation later gave rise to the Bald Eagle conglomerate of Maysville age; the nonmarine Juniata sandstone of Richmond age followed, but culmination of the movement and the end of the Ordovician were not until after deposition of the Juniata.

Similarly, the entire Upper Devonian in eastern New York and Pennsylvania is properly interpreted in connection with the Acadian orogeny, the Catskill and related facies being typical "flysch" deposits formed during the rise of a mountain belt. Successive uplifts interspersed with times of less movement seem to be indicated by the variations in coarseness of the detritus derived and transported from the mountains. These mountains reached their greatest height in the latest Devonian.[3] Great overthrusts were a part of the Taconic revolution, and Acadian intrusions on a large scale have made New Hampshire the "Granite State"; but these have not yet been tied very closely into the sedimentary time-scale.

Likewise the Laramide orogeny comprised a succession of foldings from late Cretaceous into Eocene times. Nor was the succession the same in all parts of the Laramide mountain belt. Not only did the intensity of orogenic stress vary from time to time but both the stress and the resistance of the rocks varied from place to place. It is evident that rock resistance may be overcome in one area at a certain time while in some other portions of the orogenic belt the stress-strength ratio was such that the rocks did not then give way or yielded only slightly. But at other times conditions became such that these other portions of the belt underwent folding. A major orogeny is thus an aggregate of deformations extending over a considerable length of geologic time.

For stratigraphic columns and map legends sharp dividing-lines between geologic systems are very desirable. In some cases a satisfactory division has not been difficult to find; in other instances, however, an acceptable separation can be made only after a most careful evaluation of the sedimentologic, structural, and paleontologic factors which are particularly pertinent. The "Laramie question," or where to livide the Mesozoic and Cenozoic, has perplexed geologists for half a century. In reality there was a time of transition during which profound changes took place; but structures, sediments, plants, invertebrates, and vertebrates did not all change together at the same rate, and any sharp separation of these eras is essentially a compromise for practical purposes. The Laramide revolution was a protracted succession of events.

The point to be emphasized in this discussion is that the whole duration of the Pleistocene was short compared with the span of an orogenic revolution and that the Pleistocene's high-standing continents, restricted epicontinental seas, and diversified climates suggest that it was still within the twilight between geologic periods. Notable folding of Pleistocene beds and present seismic activity in well-defined belts are in harmony with this concept.

[3] Ely Mencher, "Catskill Facies of New York State," *Bull. Geol. Soc. Amer.*, Vol. L (1939), pp. 1761–94.

REVIEWS

Minerals and Rocks. By RUSSELL D. GEORGE. New York: D. Appleton–Century Co., 1943. Pp. xviii+595; pls. 48; figs. 150. $6.00.

This book with its very numerous excellent photographs describes minerals and rocks from the point of view of the economic mineralogist. Roughly 340 pages are devoted to the description of a great number of minerals and their utilizations. Minerals are classified into three divisions: metallics, nonmetallics of industrial importance, and rock-formers. The first is subdivided according to ferrous, major nonferrous, and six other categories. The second has fuels, fertilizers, hydrosphere-atmosphere, and gems. The third has primary, secondary, and a chapter on uses. Some 123 pages are used to describe rocks (igneous, sedimentary, metamorphic) and their industrial uses.

Two introductory chapters cover (1) chemical, physical, and crystallographic properties (21 pages), and (2) genesis of ore deposits (23 pages). The section on determinative mineralogy (pp. 389–430) is devoted mainly to the blowpipe and tests for the elements. The book lacks determinative tables of any kind. There is an appendix classifying materials by uses, a short glossary, and a complete index.

The book is very attractive for one so purely practical and largely descriptive. Many mineralogists will dislike the classification of minerals, which separates zoisite from epidote (pistacite), for example. On the other hand, the classification has undeniable advantages that can be appreciated even by those in favor of the more standard form. More serious to others will be the very brief section on crystallography. Such things as axial ratios, Miller indices, and cleavage angles (in general) are omitted. There is no space for the polarizing microscope or any kind of a goniometer. There is no indication that the X-ray work of the past half-century has had any impact on mineralogy; even the formulas for the amphiboles are not modernized. Incidentally, the way the book is set up, the student would consider olivine to be an amphibole.

The work has chapters devoted to mineral fuels and to the atmosphere-hydrosphere; in this respect it is better than many recent mineralogies. However, the former is much more largely economic than it is mineralogic or petrographic. Another modern feature of the book is the chapter on the radioactive series. This would be improved by a diagram or two; also by the statement that alpha particles are helium ions rather than atoms. Similarly (p. 6), most mineral crystals grow ion by ion, and not molecule by molecule.

Professor Emeritus George has succeeded to a remarkable degree in his endeavor to "survey the whole field of earth materials." Some twenty-five years ago he wrote a similar but smaller volume for the Colorado Geological Survey. Thus the present work represents a ripe accumulation of the long life of a successful field geologist. The publishers have done a very creditable job in compressing so much typographically pleasing information into $1\frac{1}{4}$ inches on the bookshelf.

D. J. F.

A Brief Account of the Mineralogy of the Graphite Deposits of Ceylon: A Note on the Origin of the Graphite. By D. N. WADIA. (Professional Paper No. 1, "Records of Department of Mineralogy.") Colombo, 1943. Pp. 15–24; figs. 9; pl. 1.

This is one of four papers comprising the first number of the records of the recently established Department of Mineralogy. Other papers deal briefly with the rare-earth minerals of Ceylon and with three superposed peneplaines of Ceylon; a brief bibliography of Ceylon geology is included.

Graphite has been systematically mined in Ceylon for about one hundred and twenty years, mostly from pits and mines less than 100 feet deep, though one mine now is 800 feet deep and another 1,600 feet.

Four modes of occurrence in Ceylon are: (1) as disseminated flakes in metamorphic rocks—granulites, schists, and crystalline limestones; (2) as veins of nearly pure graphite; (3) in pegmatite dikes and related quartz veins; and (4) in limestones near igneous contacts. Only the veins are of commercial importance.

The rocks associated with the graphite belong mainly to the Khondalite system, which is regarded as Archean. Usually the graphite veins have remarkable continuity in direction, running in well-defined belts along the prevalent strike of the country rocks. Occasional isolated patches or pockets occur.

The graphite veins are traceable into graphite-bearing pegmatites; and these, in turn, into quartz veins. Primary minerals occurring in

the graphite veins are: scapolite, wollastonite, forsterite, tremolite, common hornblende, phlogopite, apatite, magnetite, garnet, allanite, sphene, molybdenite, and lime-feldspars. Many of these minerals are graphically intergrown, and this is interpreted as indicating simultaneous crystallization.

The flakes of graphite disseminated in some abundance through sillimanite rocks, quartz schist, and granulitic gneiss are interpreted as due to the dynamic metamorphism of nongraphitic carbon occurring in the original shales and sandstones.

The graphite veins are clearly the result of the filling of pre-existing fractures. The author attributes the formation of the graphite to the dissociation of limestones and dolomites through contact with acid intrusives, the areas of the most active graphite mining being belts of extensive granitization. The resulting CO_2 is dissociated at these high temperatures to yield graphite, the reduction being accomplished possibly by hydrogen, though the chemistry of the process is not discussed.

The dissociation of CO_2 liberated during contact metamorphism has, of course, received consideration as a possible source of graphite by most geologists interested in the genesis of this mineral but has not found much favor because of the absence of graphite from most contact metamorphic deposits. The surprising rarity of occurrence of vein deposits of graphite favors either very unusual conditions of contact metamorphism or an entirely different mechanism of graphite formation. In most contact metamorphic zones there is no mineralogical evidence of the presence of any agent capable of reducing CO_2 to CO or to C. Even where, as pointed out by B. S. Butler,[1] the oxidation of ferrous to ferric compounds in some contact zones points to a concomitant reduction of CO_2 to CO, the reduction does not seem to have gone so far as the formation of graphite. It must be admitted, however, that, where at high temperatures and pressures both CO_2 and CO are present, changes in equilibrium with declining temperature might produce graphite by the reaction $2CO \rightarrow C + CO_2$.

The most acceptable alternative hypothesis for the formation of vein graphite is that Alexander N. Winchell,[2] based on analogy with the water-gas reaction. In the water-gas process steam reacts with the carbon of coke as follows

$$C + 2H_2O \rightarrow CO_2 + 2H_2,$$

$$C + H_2O \rightarrow CO + H_2.$$

With falling temperature, equilibrium is maintained between the products of these reactions by the further reaction, $2CO \rightarrow C + CO_2$. While in water-gas plants the carbon deposited by this reaction is nongraphitic, Winchell believes that under natural conditions of high temperatures and pressures graphite would be produced.

Under Winchell's hypothesis, water at high temperatures and pressures becomes an agent for the oxidation of older graphite (usually as flakes in meta-sedimentary schists) to CO and CO_2. In these compounds the carbon may be transported, to be redeposited as elemental graphite. No reducing agent is necessary for the deposition of graphite by this mechanism but the presence of older carbonaceous deposits is essential; and these are present both in Ceylon and at Dillon, Montana.

E. S. BASTIN

[1] "The High Ferric Oxide Content of Limestone Contact Zones" *Econ. Geol.*, Vol. XVIII (1923), pp. 398–404.

[2] "A Theory for the Origin of Graphite as Exemplified in the Graphite Deposit near Dillon, Montana," *Econ. Geol.*, Vol. VI (1911), pp. 218–30.

Archaeological Researches in the Northern Great Basin. By L. S. CRESSMAN and COLLABORATORS. (Carnegie Institution of Washington Publication 538 [Washington, 1942].) Pp. xvii+158; figs. 102. $3.00, paper; $4.00 cloth.

This volume, chiefly in the field of archeology, has resulted from the collaboration of various specialists with help from workers in related sciences. A study of the relationship of the culture of the caves and lake beds of the northern Great Basin (mostly in south-central Oregon) to the early cultures of the Southwest is the main theme. Geology has contributed some evidence on the time relations. The Paisley Five-Mile Point Cave has revealed occupation levels separated by a stratum of pumice in each case; and this pumice has been identified by Howel Williams as having come from Mount Mazama, not less than 5,000 years ago. Excavations reported in this volume have definitely established the association of man with an extinct fauna—*Equus*, camel, and some type of proboscidian.

R. T. C.